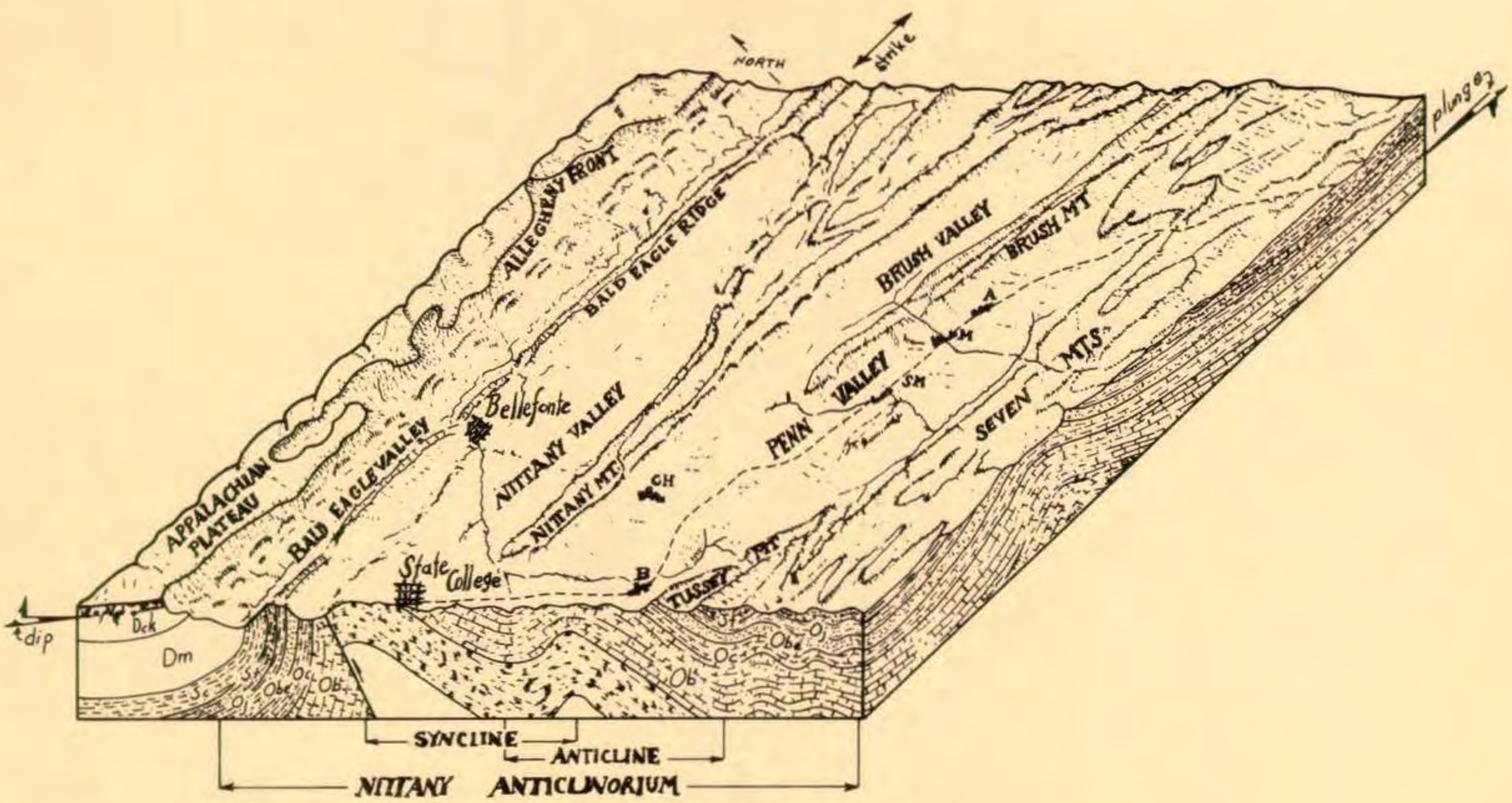
GUIDEBOOK

50th. Annual Field Conference Of Pennsylvania Geologists

CENTRAL PENNSYLVANIA GEOLOGY REVISITED



October 4, 5, and 6, 1985
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CENTRAL PENNSYLVANIA GEOLOGY REVISITED

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FIELD TRIP #1

A MID-SILURIAN CORAL-BRYOZOAN REEF IN CENTRAL PENNSYLVANIA

by
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Shirley S. Fonda, Anne B. Lutz,
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Significance and Purpose

Central Pennsylvania lies just beyond the borders of the Great Lakes district, renowned for its development as one of the world's earliest major coral reef provinces. Nonetheless, Silurian rocks in our state also contain a few small bioherms, which apparently represent scattered outliers from those major Midwestern reef tracts (Cuffey and Davidheiser, 1979; Inners, 1984).

One in particular among our local bioherms is of wider interest scientifically, for two reasons. First, unlike the Great Lakes Silurian reefs, this Pennsylvania one--near Lock Haven (Clinton County; Fig. 1)--is undolomitized and hence provides a good view of the organisms and sediments involved in reefs at that time. Second, bryozoans participated significantly in the construction of the Lock Haven reef, a rather unusual and hence intriguing situation, because reef-building represents a paleoecologic extreme for that phylum, and because bryozoan-built structures are exotic variants among bioherms through geologic time (Cuffey, 1977, 1985).

Consequently, we wish to showcase the Lock Haven reef here, so that its unique features and potential contributions can be appreciated both by Pennsylvania geologists and also by reef- and bryozoan-oriented scientists in general. Our chapter will therefore serve for overall summary, this conference field trip, and future self-guided individual visits. After briefly outlining the overall setting, we will discuss implications of our studies of the Lock Haven reef (Cuffey and Davidheiser, 1979; Davidheiser, 1980; Davidheiser and Cuffey, 1981; Cuffey, 1985), and then present its observable characteristics in the format of several field-trip stops.

Initially cut into years ago by the road (then dirt or gravel) running eastward from Castanea, the Lock Haven reef remained rather poorly exposed until mid-1972. Then, the extremely heavy rains from Hurricane Agnes provoked much slumping along the road cuts at the foot of Bald Eagle Mountain. These slumps, combined with subsequent road repairs and reconstruction, improved the reef exposures significantly for a few years. Recently, however, weathering and continued slumping are again obscuring and covering parts of the outcrop.

In the years spent examining the Lock Haven reef, we have been assisted by many individuals; we wish to particularly thank L. Brant, W. Bruck, G. Burgess, T. Davies, R. Dunay, J. Head, J. Jobling, S. Krajewski, P. Kremer, R. Lanning, R. Lowright, J. Malkames, C. McKee, G. Mertz, J. Morrison, G. Moser, G. Newton, M. Podwysocki, J. Shultz, D. Siegel, J. Swinehart, A. Thorn, K. Wilson, and M. Zeigler.

Overall Setting

In current terminology (Cuffey, 1985), the Lock Haven reef consists of a favositid crust-mound center, surrounded by cladoporid and trepostome frame-thickets. The exposed surface of the Lock Haven reef displays lateral variations in both rock and fossil characteristics, variations which can be expressed well by grouping those materials into several lithotypes, each interpretable as having formed within a different environmental zone. Commencing at the outer edge of the reef (i.e., at the eastern end of the currently available outcrops), and progressing inward (westward) toward the central or thickest part of the reef, the sequence of lithotypes--and inferred equivalent ecozones--is platy shale and limestone (off-reef mud bottom), massive calcarenite (surrounding sand apron), trepostome branch coquina (deeper reef flank), cladoporid branch coquina (shallower reef flank), favositid reef-rock (reef crest or core), and stromatolitic rubbly limestone (reef cap) Fig. 1; Table 1).

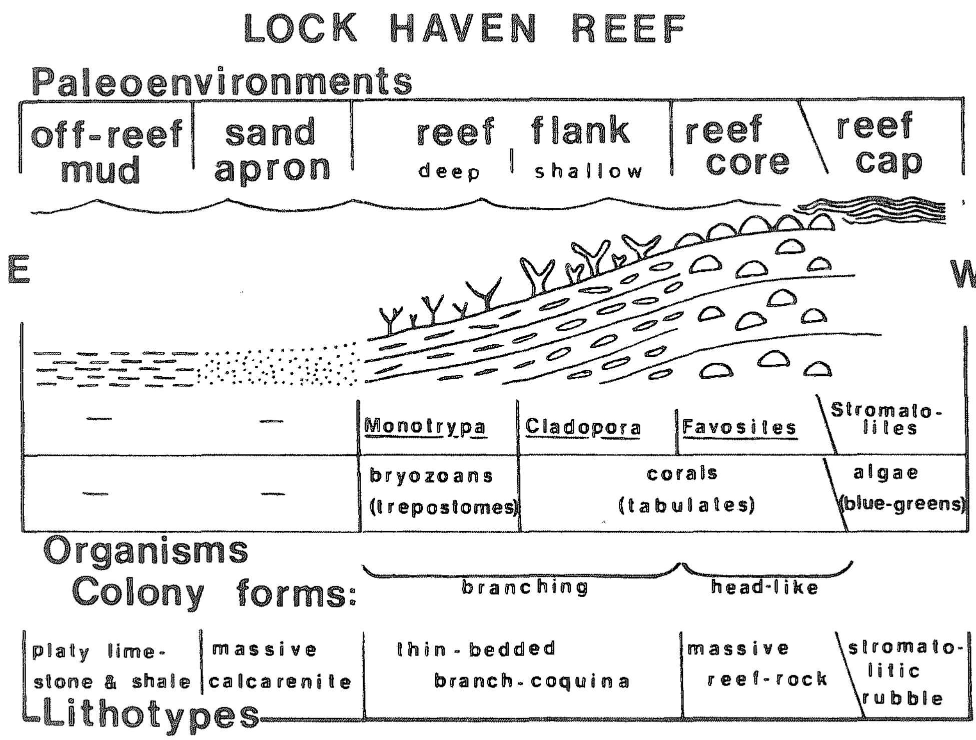

Figure 1. Ecozonation inferred for Lock Haven reef; also compare Table 1.

The Lock Haven reef is a thick carbonate, surrounded by thinly interbedded shales and limestones, which represent the Rochester Shale overlain by the lower limestone-and-shale member of the McKenzie Formation (or equivalent portions of the more broadly conceived Mifflintown Formation of recent usage). These strata are of Middle Silurian age (specifically middle Wenlockian). The reef mass can be correlated with particular biozones (as discussed later under Stop 6); it appears to be founded on approximately the Rochester-McKenzie boundary and projects upward (hence as a knob of "Rochester") into the basal McKenzie.

Regional stratigraphic correlations tie the Rochester-McKenzie contact into the New York standard section just above the Clinton-Lockport boundary (Berry and Boucot, 1970, at the level of the Gasport Dolomite, basal Lockport). The Gasport, in westernmost New York state and adjacent Ontario, contains well-developed small reefs (Crowley, 1973; Crowley and Poore, 1974), partly resembling the Lock Haven reef with which they are contemporaneous. (Lichenaliid-bryozoan reefs described from the basal Rochester or uppermost Irondequoit in that area by Sarle, 1901, are somewhat older; Hewitt and Cuffey, 1985; Brett, 1982, 1983a, 1983b).

Moreover, as a mid-Silurian reef, the Lock Haven deposit is at least roughly correlative with the vast Great Lakes Silurian reef complex. Several different times of reef initiation and growth (reef "generations") can be distinguished within that reef complex (Shaver, 1977; Shaver and others, 1978). One early, short-lived generation, early in middle Wenlockian time, consists of the Gasport reefs, and hence also the Lock Haven reef as a contemporaneous southeastern outlier.

During the Middle Silurian, North America was in an equatorial position, mostly shallowly submerged, and dominated by carbonate sedimentation, much like the modern Bahama Banks. Still isolated from most other land masses, the continent was being approached closely by one other, the Baltic-shield craton (Ziegler and others, 1977; Dott and Batten, 1976). A large region, centered on the Michigan Basin but extending outward far across the surrounding shallow shelves, developed a vast reef complex. The edge of this reefal sea graded eastward into a shallow muddy bottom perhaps slightly deeper or basinal, which in turn abutted against mountainous islands along the eastern margin of the continent. A few small patch reefs developed in that mud-bottomed Appalachian sea, and one of them is now preserved and exposed as the Lock Haven reef.

TABLE 1. Summary of observed and inferred features of the Lock Haven reef

PRESERVED FACIES SUMMARY NAME (stop numbers)	OFF-REEF SHALE (5)	CALCARENITE APRON (4)	TREPOSTOME FLANK (3)	CLADOPORID FLANK (2)	FAVOSITID CORE (1)	REEF CAP (7,8,9)
environment	open level bottom		reef mound or bioherm			
ecozone	off-reef mud bottom	circum-reef sand apron	reef flank or slope: deep flank	shallow flank	reef top: reef crest or core	reef flat
habitat	barren mud	barren sand	bryozoan thickets	coral thickets	coral heads	algal domes
depth	deep (but see text)		moderately deep	moderately shallow	very shallow	intertidal
dominant phylum or class	--	--	bryozoans	corals		algae
dominant order	--	--	trepostomes	tabulates		stromatolites
dominant genus and species	--	--	<u>Monotrypa benjamini</u>	<u>Cladopora seriata</u>	<u>Favosites niagarensis</u>	<u>Collenia</u> sp.
dominant growth form	--	--	branches		heads or domes	
general rock type	shale	limestone				
particular lithotype	platy shale	massive calcarenite	branch coquina		massive biolithite	stromatolitic cap
specific reef-rock or carbonate-rock types	mudstone marlstone, micstone	grainstone	bafflestone, rudstone, floatstone	rudstone	globstone	cruststone, bindstone, rudstone
bedding characteristics	shaly	massive	thin-bedded		massive	massive laminated

Reef Recognition

Recognition of the Lock Haven deposit as a fossil reef results from several detailed observational comparisons, rather than its geometric form (because that is so poorly exposed).

First, the suite of organisms preserved there abundantly--the coral heads, and coral and bryozoan branches--are typical of reef deposits in general, rather than of level bottoms. Indeed, this particular fact was what triggered Hoskins' (1964) original recognition of the reefal nature of this deposit.

Second, this assemblage does not occur as a coherent or consistent combination elsewhere in the enclosing formations, which yield a quite different brachiopod-ostracod fauna instead. Such sharp faunal contrast is characteristic of reef versus off-reef habitats. Patches of favositid heads are found elsewhere, but not the entire assemblage of favositids, cladoporids, and trepostomes.

Third, spatial or distributional variations within the coral-bryozoan assemblage here are the same as observed in other fossil and modern reef complexes, where they are clearly an ecozonation.

Fourth, the sedimentary matrix associated with each of the major coral and bryozoan types is typical of the organism-sediment combinations seen in ancient and modern reefs.

Fifth, fossil fragments from the zones interpreted as topographically higher occur occasionally in the matrix of zones thought to be lower, but not vice versa. This suggests downward transport along the flanks of a higher-standing, hence typically reefal, carbonate mass.

Sixth, the Lock Haven materials match analogous portions of the well-exposed Silurian reefs in the Great Lakes district. Detailed examination of a number of those reefs in Ohio, Indiana, and Illinois provided much comparative data helpful in precisely interpreting the Lock Haven reef after its exposures improved in 1972. (Hurricane Agnes that year triggered extensive slumping which temporarily much better exposed the reef than before or since.)

Seventh, the Lock Haven materials correspond nicely to comparable sections of modern Caribbean reefs on which we have dived extensively. This experience with modern counterparts proved especially useful in illustrating this paper (Figs. 4-5 of field guide).

Finally, Swartz (1970, pers. commun.; 1946, 1939, 1935b) had briefly noted "coralline lenses," with abundant *Cladopora multipora* [or *seriata*] and *Favosites niagarensis*, in the lower McKenzie near Lock Haven. He interpreted them as Lockport coral-rich limestone tongues from the north interfingering with McKenzie ostracod-bearing shales to the south, but did not label them as reefs. Most probably, his not stating the reefal character of these deposits was a function of his times, when reef geology was in an embryonic state, prior to the extensive biogeologic studies resulting from naval actions in the Pacific years later.

Reef Growth History

Only the outermost edge of the foundation underlying the Lock Haven reef is currently exposed; it is the platy shale and limestone typical of the off-reef habitat. The stratigraphic position of the reef mass suggests that such rocks extend underneath the entire reef. The lowest ecozone within the mound is the surrounding calcarenite sand apron; it is conceivable--though not demonstrated yet--that this rock type also goes across under the whole reef, as its basal zone. Apparently, therefore, the Lock Haven bioherm was initiated upon either a mud or sand bottom, mixed terrigenous and calcareous in composition, and likely soft or shifting (unstable, in either case). Such a foundation is not typical for modern coral reefs (though may possibly be approached by the rare Bahamian bryozoan reefs; Cuffey and others, 1977), but is encountered in many Paleozoic reefs. Various Great Lakes Silurian reefs, for example, were initiated upon terrigenous mud, carbonate mud, or carbonate sand (but not hardgrounds, so far as reported).

Once the reef foundation was in place, whether mud or sand, organic growth activities by colonial invertebrates became the predominant processes. Low on the reef flank, clumps of branching trepostomes (Monotrypa benjamini) trapped or baffled carbonate sands. Several such colonies are now visible, standing in place, still largely embedded in the calcarenitic matrix (Fig. 4J of field guide). Such sediment-trapping clumps may well have been the first reef organisms to occupy the barren sand patch on the overall muddy bottom. Higher on the reef flank, branching cladoporids (Cladopora seriata) apparently continued with the sediment-baffling role. In addition, throughout the flank, broken trepostome and cladoporid branches accumulated voluminously as skeletal sediment. However, outside of the branching colonies themselves (Fig. 4P of field guide), none appear to have formed an actual interlocking open framework here. In contrast, the head-like favositids (Favosites niagarensis) built a massive, dense, largely solid, lump-upon-lump reef frame in the reef core or crest, and also contributed some large loose boulders as reef-top loose rubble. Shells of other groups--particularly gastropods, calcispheres, and ostracods--added more skeletal sediment, especially around the foot of the reef flank. No bryozoans appear in the cryptic hidden-encrusting role seen elsewhere in some Silurian (Scoffin, 1972; Spjeldnaes, 1975) and most modern (Cuffey, 1977) reefs.

Zones within certain fossil reefs elsewhere have been interpreted as successive seral stages, from pioneer to climax communities. In contrast, the various ecologic zones seen within the Lock Haven reef appear to have been synchronously coexisting, rather than chronologically successive. As a result, except for possibly distinguishing an earlier normal-marine reefal stage (with several simultaneous ecozones) and a possibly slightly later intertidal stromatolitic-cap stage (reminiscent of several Michigan Basin reefs; Mesolella and others, 1974; Huh and others, 1977), little can be inferred concerning growth stages within that bioherm. Moreover, even if the Lock Haven ecozones were successional, difficulties would still remain in applying published developmental patterns to this reef. Mid-Ordovician reefs exhibit pioneer colonization and stabilization stages, followed by biotic diversification, and eventually climaxed with domination by a particular reef-building group (Alberstadt and others, 1974); at Lock Haven, however, each zone is occupied by only one major species, with no visible diversity changes from one to another. Great Lakes mid-Silurian reefs are described as growing from deep quiet-water pioneer, upward through semi-rough, into shallow rough-water climax stages (Lowenstam, 1957; Nicol, 1962). The reefs exhibiting this seral succession are hundreds of feet thick; in contrast, the Lock Haven reef is so thin or low that each zone could have been only barely different in turbulence conditions from those preceding and following it.

Although regional studies indicate "shallow" waters for the Lock Haven reef, the actual reef form and zonation may imply a more precise figure for water depth there. In particular, stratigraphic relief over the mound can be interpreted as also being the during-life depositional relief; thus, a minimum of about 30 feet (10 meters) water depth is implied. Moreover, the algal stromatolite cap was probably intertidal; if this cap grew at the same time as (or very shortly after) the reef core which it partly covers, then mean sea level would have been right at the top of the mound, and hence the actual water depth immediately surrounding the reef would have been on the order of 25 or 30 feet, at least at the very end of reef growth.

Another intriguing but more complex possibility is suggested by recent studies of temporary low sea-level stands during Silurian time (Dennison and Head, 1975). Most of this time, the Appalachian sea may have been at a relatively constant depth, shallow but not extremely so--perhaps on the order of as much as 200 feet (60 meters). At several points in time, sea level dropped throughout the region (possibly due to world-wide eustatic changes or plate-tectonic movements, or to regional tectonic warpings) for a geologically short moment and then rapidly recovered to its "normal" level; some of the drops were apparently slight, but others may have virtually emptied the basin. Perhaps, the initial Lock Haven trepostome thickets baffling calcarenite sand grew under those deeper waters, and by coincidence sea level began to drop rapidly just after their establishment. As the water column shallowed to intermediate depth, the trepostomes were replaced by cladoporids, in turn giving way to favositids as shallowing continued; finally, the reef top became intertidally exposed to develop the stromatolites capping the reef. If rates of reef growth and sea-level change were comparable to those seen in the post-glacial Pleistocene-Holocene transition, the two processes might well have nearly balanced, thereby producing only a relatively thin reefal deposit during each rapidly shallowing depth interval.

The principal problem with this possible interpretation of Lock Haven reef history is that the inferred sea-level lowerings do not coincide with the geologic time of reef growth; lowered sea level is indicated by the Keefer sandstones well below, and next by the Rabble Run red beds well above, the horizon of the Lock Haven reef (Dennison and Head, 1975). Moreover, although the outcrops do not permit full disclosure, the Lock Haven reef rocks do not appear to exhibit any features suggesting episodes of exposure and drying of the reef mass (there are, for instance, no Barbados-type caliche horizons cutting across coral heads within the reef core). Therefore, the reef probably remained permanently or continuously submerged during its growth and development (unlike some of the Great Lakes Silurian reefs).

After full reef development, however, the stromatolitic cap indicates intertidal conditions and hence intermittent exposure (probably briefly, until resumed mud accumulation off-reef buried the reef mass early in McKenzie time, at most only a few thousand years following Rochester time). No obvious erosional surface or truncation was seen in the available outcrops; hence, while exposure may have contributed to reef termination, scouring does not seem to have been involved much (although the breccia or rubble in part of the reef cap suggests some erosion).

Overall duration of Lock Haven reef growth was probably quite short geologically, as implied both by the thinness of the reef mound (only 30 feet thick) and by its being embedded entirely within the Rochester-McKenzie shale sequence (a small fraction of the total Middle Silurian interval). Moreover, the reef generation to which the Lock Haven (and the Gasport) reefs belong flourished entirely within the early part of mid-Wenlockian time, again a relatively short interval within the Silurian period. Thus, the life span of the Lock Haven bioherm seems comparable to that of many Holocene patch reefs, whose individual durations have been determined as on the order of a few hundred or very few thousands of years.

Comparison with Great Lakes Silurian Reefs

As noted previously, the Lock Haven reef can be regarded as an eastern outlier of the vast Silurian reef complex of the Great Lakes district, and hence can be profitably compared with well-developed reefs in that region (Briggs and others, 1978; Crowley, 1973; Droste and Shaver, 1977, 1983; Lowenstam, 1957; Shaver, 1977; Shaver and others, 1978). Indeed, many aspects of the Lock Haven reef became fully understood only after careful field examination of several of the Great Lakes reefs.

An obvious difference is the geographically isolated position of the Lock Haven reef, far away from contemporaneous reef tracts. The Lock Haven reef sat out on a sea floor dominated by terrigenous muds, while the Great Lakes reefs were surrounded by carbonate-sediment bottoms.

Moreover, the Lock Haven reef is relatively thin or low-standing, and small and possibly equidimensional in plan view, as are many similar patch reefs in the Great Lakes area. Other Great Lakes reefs are much thicker, up to several hundred feet stratigraphically, and extended laterally as elongate barrier reefs.

Another conspicuous contrast is the predominantly limestone composition of the Lock Haven reef, versus the thoroughly dolomitized character of the Great Lakes reefs.

Both the Pennsylvania and Midwestern bioherms exhibit well-developed ecologic zonations. The Lock Haven reef, and many Great Lakes reefs, have a deeper zone characterized by branching trepostomes and/or cladoporids, functioning as sediment trappers and as sediment formers. Shallower zones in these are dominated by massive frame-building heads, although the Lock Haven head-like colonies are favositid corals, while the Great Lakes ones are predominantly stromatoporoids with only minor favositids. And, numerous Michigan Basin reefs--like the Lock Haven reef--are capped by stromatolitic deposits, probably representing at least brief intertidal exposure at the end of reef growth. Zonations in the Lock Haven and many smaller Great Lakes reefs appear to have been contemporaneously coexisting ecologic zones on the surfaces of relatively short-lived reef mounds; in addition, zonation in many larger Midwestern reefs reflects successional seral stages in comparatively long-duration reef growth.

The organic communities building and inhabiting the Lock Haven and Great Lakes reefs, although comparable, are somewhat different. The same major groups, for the most part, are found in each. However, stromatoporoids, pelmatozoans or crinoids, and fenestrate bryozoans are missing from the Lock Haven reef but are present (and often abundant) on Midwestern Silurian reefs. Additionally, the Lock Haven assemblage is less diversified at lower taxonomic levels; among the tabulate corals, for example, only favositids and cladoporids occupied the reef mound, while those two plus halysitids, alveolitids, and others flourished on the Great Lakes reefs.

Other Pennsylvania Silurian Bioherms

Another small bioherm occurs at roughly the same horizon as the Lock Haven reef, about 25 miles further east near Allenwood (Inners, 1984). It is composed of Favosites heads, so that it appears comparable to the reef core at Lock Haven; however, it lacks the surrounding broken-branch rubble and calcarenite apron of the latter. It thus resembles the small satellite reef (Stop 10) adjacent to the Lock Haven reef (Fig. 5T of field guide).

At Lakemont, on the south side of Altoona, is a limestone lens (Swartz, 1935a, 1939), also approximately at the Rochester-McKenzie boundary. However, field examination of the highway cut exposure there shows mostly stromatolitic masses interbedded with shales and limestones more reminiscent of tidal-flat than reefal deposits; a couple of rolled favositid cobbles recovered from the base of the lens suggest possible nearby coral patches. Indeed, current excavations several hundred feet to the northeast are encountering scattered favositid heads heavily encrusted with algal stromatolites, materials which may occur in this stratigraphic interval relatively widely across central Pennsylvania (A. L. Guber, 1984, pers. commun.). If confirmed by further investigation, such would indicate scattered coral patches, some of which grew into small reef mounds like that at Allenwood, and a few into larger patch reefs like that at Lock Haven.

According to Swartz (1970, pers. commun.), three or four miles west of the Lock Haven reef may be another bioherm, 30-40 feet thick, behind the funeral home and agricultural office in the village of Mill Hall, but possibly covered or removed by recent highway construction (according to local residents). Several field visits have so far failed to yield anything other than typical non-reefal Rochester-McKenzie limestone slabs. Additionally, Swartz (1935b) referred to "coralline deposits" near Williamsport as well as Lock Haven, but no exact locations were given.

At a higher horizon (the Keyser Limestone), a small bioherm exposed in Altoona (Brezinski and Kertis, 1982) consists of a lower foliaceous bryozoan lettucestone zone overlain by an upper stromatoporoid reef-rock cap, and so presents quite a different combination of taxa than does the Lock Haven reef.

Finally, a few patch reefs have been noted in the West Virginia Silurian, and some at least have cladoporid zones resembling that at Lock Haven (Patchen and Smosna, 1975; Smosna and Warshauer, 1978, 1979, 1983).

FIELD TRIP #1

GUIDE TO THE LOCK HAVEN REEF

by
Roger J. Cuffey, Carolyn E. Davidheiser, Shirley S. Fonda,
Anne B. Lutz, Laurie S. Zimmerman, and Barbara B. Skerky

DRIVE TO LOCK HAVEN ON DIVIDED HIGHWAY U.S. 220; TAKE EXIT MARKED "LOCK HAVEN, PENNA. HWY. 120 WEST; CASTANEA, LOCK HAVEN UNIVERSITY"; TURN SOUTH (TOWARD MOUNTAIN, AWAY FROM CITY); IN 0.1 MILE, TURN LEFT (EAST) AT T-JUNCTION; 0.2 MILE FURTHER, TURN RIGHT (SOUTH) AT T-JUNCTION BESIDE CASTANEA FIRE COMPANY #1 BUILDING.

DRIVE SOUTH (TOWARD MOUNTAIN) FOR 0.2 MILE, ACROSS BALD EAGLE CREEK, INTO THE VILLAGE OF CASTANEA (WHERE THE ROAD BECOMES LOGAN AVENUE); TURN LEFT (EAST) ON BROWN STREET; IN 0.1 MILE, TURN RIGHT (SOUTH) ON McELHATTAN AVENUE; 0.1 MILE FURTHER, TURN LEFT (EAST) ON EAST KELLER STREET; PROCEED EASTWARD FOR 0.9 MILE TO PARKING PULLOUT AREA ON RIGHT (SOUTH) SIDE OF ROAD. PARK AND WALK BACK TOWARD CASTANEA TO SEE THE REEF EXPOSURES IN THE LOW ROADCUTS ALONG THE SOUTH SIDE OF THE ROAD (FIGS. 1 AND 2).

FROM THE CENTER OF THE PARKING AREA, WALK WESTWARD (BACK TOWARD CASTANEA) 490 FEET (149 METERS) TO PROMINENT EXPOSURE OF RUBBLY-WEATHERING MASSIVE LIMESTONE (STOP 1) IMMEDIATELY EAST OF TELEPHONE POLE NUMBERED "09209 N34998."

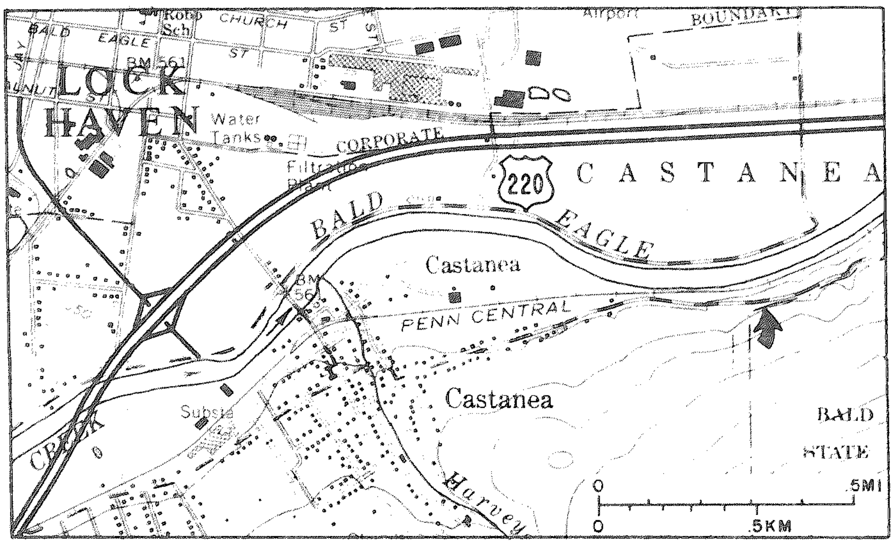

Figure 1. Location of Lock Haven reef (arrow); base modified from Lock Haven and Mill Hall (Pennsylvania) 7-1/2' topographic quadrangles, U.S. Geological Survey.

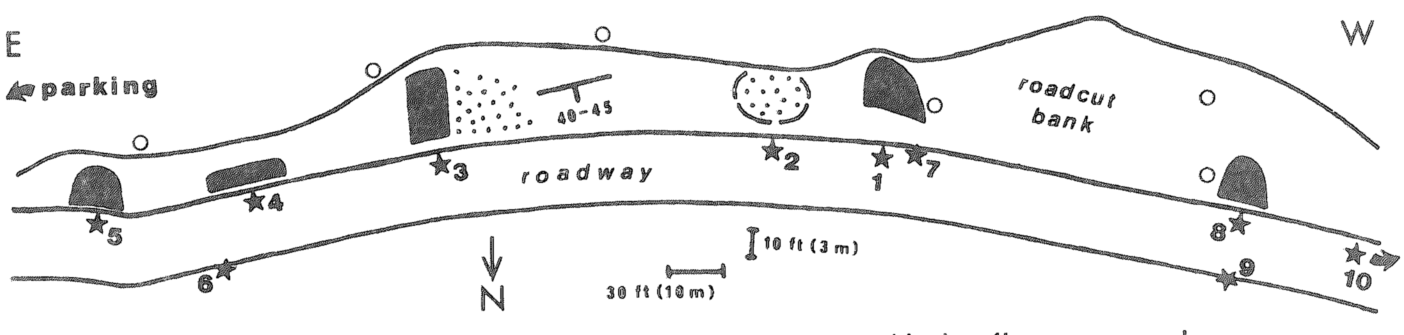

Figure 2. Detailed map of Lock Haven reef (rock exposures, black; float-strewn slopes, stippled); stops indicated by numbered stars; open circles are bases of telephone poles; strike-and-dip symbol indicates average for reef materials, mostly N75-80°E and 40-45°N; note exaggeration of cross-roadway versus along-roadway distances.

Stop 1: Favositid Reef-Core

This exposure shows the center of the Lock Haven reef deposit. That center is massive reef-rock or biolithite, made up of favositid coral heads, and represents the ancient reef crest (Fig. 5G-M).

The rock here is globstone, medium-gray, tough, hard, crystalline, and composed of in-place heads of tabulate corals (Favosites niagarensis, up to a foot in diameter). This lithotype weathers light brown, and bouldery, nodular, or rubbly appearing (due to weathering around the Favosites heads). The coral heads are separated by minor amounts of in-filling micrite mud and spar cement; occasional calcispheres and brachiopod fragments are embedded therein as well.

This massive favositid reef-rock, during life, was probably a rocky or rubbly surface almost completely covered with head-like coral colonies (Favosites niagarensis), and probably very shallow and wave-swept. The head-like brain corals (Diploria spp.) presently covering reef-knoll tops on the Bermuda outer reefs furnish a modern analogue to this ecozone (Fig. 5H).

WALK EASTWARD (BACK TOWARD PARK AREA) 55 FEET (17 METERS) TO FLOAT-STREWN SLOPE (STOP 2).

Stop 2: Cladoporid Reef-Flank

Although now largely covered, the bedrock here is indicated by the numerous float slabs of cladoporid branch coquina, which represent the original shallow reef flank (Fig. 5A-F).

Thin-bedded, dark-gray to medium-brown, tough to friable, densely fossiliferous limestones (mostly rudstones, a few floatstones; originally bafflestones, but not branchstones) comprise this lithotype, characterized by closely packed, broken, narrow branches lying parallel to bedding and making up the great bulk of the rock volume. The matrix in which the branch fragments are embedded appears aphanitic in hand specimen, but is varied in thin-section--partly clay-mineral mud, partly micrite, and partly finely crystalline spar cement. The branches are mud-filled, moderately finely tubular, moderately small apertured, cladoporid tabulate corals (Cladopora seriata). Local pockets or clusters of small brachiopods are scattered through the cladoporid coquinas. Near the top of this ecozone (adjacent to the reef core), a few overturned and fragmentary favositid heads intermingle with the cladoporid branches.

The cladoporid-branch coquina was the shallow, probably still turbulent portion of the old reef flank or side, gently sloping away from the crest into somewhat deeper water. Densely covered with the projecting tips of thinly branching colonies (Cladopora seriata), this ecozone must have closely resembled the dense branching Acropora thickets around the edge of the Heron Island (Australia) platform reef, or the tangled thickets of Acropora cervicornis on Caribbean reefs (Fig. 5B). Locally, clusters of small brachiopods nestled down among the cladoporid branches.

WALK EASTWARD AGAIN, 165 FEET (50 METERS) TO HIGH EXPOSURE OF SMOOTH-SURFACED MASSIVE LIMESTONE (STOP 3).

Stop 3: Trepostome Reef-Flank

The exposed limestone here is a trepostome bafflestone, with several large in-place bryozoan colonies embedded in the carbonate sands. To the right (west) of this exposure, the bedrock is mostly covered, but float blocks of trepostome floatstone and rudstone indicate that these branch coquinas extend over toward the cladoporid coquina seen at the previous stop. These bryozoan coquinas represent the ancient deep reef flank ecozone (Fig. 4I-R).

These rocks overall appear identical to the cladoporid coquina described at the preceding stop, except for the characteristics of the branches themselves. Here, they are spar-filled, very finely tubular, very small apertured branches identified as Monotrypa benjamini, a trepostome bryozoan. Near the base of this ecozone (adjacent to the massive calcarenite sand apron), the matrix is more calcarenitic, with many gastropod, ostracod, and calcisphere fragments, and a few branching-trepostome clumps or thickets (up to 3 feet or 1 meter across) still standing in place; these, again, are Monotrypa benjamini.

The trepostome-branch coquina also probably looked very much like the preceding ecozone when alive, but in somewhat deeper and probably a bit quieter waters. Finely branched thickets of Acropora prolifera in the reefs off St. Croix (Fig. 5K) furnish a modern analogue.

WALK EASTWARD AGAIN, 95 FEET (29 METERS) TO LOW EXPOSURE OF SMOOTH-SURFACED MASSIVE LIMESTONE (STOP 4).

Stop 4: Circum-Reef Calcarenite Apron

Superficially much like the limestone seen at the previous stop, the bedrock here, however, contains few or no bryozoan colonies and fragments. It is thus a massive calcarenite, a lithified carbonate sand originally deposited as a barren apron surrounding the reef mass (Fig. 4E-H).

This rock type is specifically a grainstone, dark-gray (weathering dark brown), massive, unfossiliferous, homogenous, hard, tough, and smooth-surfaced to vaguely ripple-marked. Freshly broken surfaces appear coarsely crystalline or obscurely clastic, with relatively uniform grain size; weathered surfaces become sandy textured. Nearer to the reef mass proper, occasional small trepostome-branch fragments may be found embedded in this calcarenite. A rather different appearance is presented by thin-sections of this lithotype, being composed of much fossil debris surrounded by subordinate matrix material. The fossils are small, broken, often recrystallized fragments of abundant gastropods, calcispheres (of uncertain affinities, but possibly algal or protozoan), and ostracods, accompanied by a few brachiopod or trilobite fragments. Most of the embedding matrix is clear crystalline spar, largely cement, but locally recrystallizing or even cross-cutting in origin.

The massive calcarenite represents the bottom flattening out at the foot of the reef flank. This area immediately surrounding the reef mass was level-bottom loose carbonate sand, open or barren (i.e., lacking any colonies growing in-place here), and locally ripple-marked. Many modern reefs--such as patch reefs within the Florida reef tract and on the shallow shelf east of Joulters Cays on the Great Bahama Bank--are surrounded by such barren sand aprons (Fig. 4F).

WALK EASTWARD AGAIN, 85 FEET (26 METERS) TO SIZEABLE EXPOSURE OF PLATY SHALE AND SHALY LIMESTONE (STOP 5).

Stop 5: Off-Reef Shale

The shaly bedrock here is not part of the Lock Haven reef itself, but rather represents the ancient deep off-reef muddy bottoms above which the reef grew (Fig. 4A-D).

The platy off-reef lithotype consists of calcareous shale, shaly limestone, and thin-bedded limestone (i.e., mudstones, marlstones, and micstones), all intergrading continuously with one another; they are medium brown to medium gray brown when fresh, but weather light gray brown. Evenly bedded, not rippled at all, the beds are hard, brittle, and platy. Their texture is aphanitic and argillaceous; they weather to a finely muddy surface. Largely unfossiliferous, these rocks exhibit only a few horizontal trails or burrows on some bedding planes. In thin-section, these platy strata consist partly of brownish, opaque, clay-mineral mud, and partly of grayish, homogeneous to clotted (pelleted) micrite, all thinly interbedded. No detrital quartz sand grains were seen. A few small fossil shell fragments appear in the sections, a few gastropods and occasional ostracods, but no bryozoans or corals.

These rocks represent off-reef mud bottoms. They were possibly barren but possibly vegetated (presumably with soft-bodied algae not now fossilized), certainly lacking colonies, but locally strewn with small shells, burrowed, level, and apparently at the same depth as the adjacent sand apron. Carbonate-mud bottoms under 12-15 feet (4-5 meters) of water on the Bahama Banks provide a suitable modern counterpart (Fig. 4B).

STEP BACK NORTHWARD, ACROSS THE ROAD (BEWARE OF ONCOMING TRAFFIC!); WALK WESTWARD AGAIN (BACK TOWARD STOP 1), BUT FOR ONLY 60 FEET (20 METERS); STOP AND FACE WESTWARD TO VIEW SPATIAL RELATIONSHIPS OF STRATA SO FAR EXAMINED (STOP 6).

Stop 6: Overview of Reef Form

The Lock Haven reef is poorly exposed, and therefore its overall mounded form is difficult to discern in the field. However, the best opportunity for doing so is from this spot, after the observer has examined each of the facies preserved at the previous stops.

Standing here, at the eastern end of the reef exposures and sighting back (westward, along strike) at the thickest portion of the reef, one recognizes about 30 feet (10 meters) distance stratigraphically between the base and top of the reefal limestone beds. Behind his back (further eastward), the winding road in places cuts through that same 30-foot interval, but there it consists entirely of shaly beds without evidence of reefal involvement. Moreover, some of the lower limestones--especially those at Stop 3 containing in-place branching trepostome clumps--exhibit a mounded upper surface (over a thickened portion of that bed, Fig. 4L, so that the mound is clearly not the upwarped surface of a small anticline). This mounding appears to slope gently upward toward the stratigraphically highest favositid reef bed; if the two are mentally connected, they form an upper surface (for the limestones) which is inclined slightly (perhaps only 5°) to the overall dip of the shaly sequence. It is, moreover, this limestone surface off of which the overlying shaly beds slumped after hurricane rainfall in 1972, thereby exposing the surface rather cleanly so that lateral variations within the limestones could be readily examined.

These observations suggest that the limestones exposed comprise one side or half of a lenticular or mound-like mass, still largely buried within the mountainside, but only thinly covered along the road by shales not significantly enough interfingered with the carbonates to prevent slumping off of that cover as a whole. By analogy with modern reefs similarly isolated from other contemporaneous reefs, a mound-like form of roughly circular plan view would be a reasonably likely shape for the Lock Haven reef; its dimensions would then be 30 feet (10 meters) high by approximately 1000 feet (300 meters) across. The lack of interfingered shale along the limestone surface implies lack of simultaneous reef and shale deposition. In other words, the reef mass probably stood up as a knob--perhaps nearly as high as its 30-foot (10 meter) thickness--above the surrounding mud-bottomed sea floor for a short time geologically. Later on, the reef became buried by further mud accumulation.

Another implication of these geometric relationships is that the exposed limestone surface approximates the ancient living surface of the reef mound, so that lateral variations exposed now within the limestone represent contemporaneous ecologic zones (ecozones) coexisting down the flank of the reef mass during its life. Again by analogy with living reefs of similar form, those ecozones probably had coexisted for some time prior to the instant represented by the exposed surface of growth and deposition; hence, each ecologic zone visible at the surface most likely extends below in lateral facies relationships like those typically seen in better exposed fossil reefs (Fig. 3). Future drilling or deeper cut exposures would, however, be necessary to fully test this interpretation. An alternative possibility would be to interpret the various limestone types as being flat layers resting one on top of the other; however, reefs seldom, if ever, grow in such fashion, so that this seems quite unlikely.

In addition to this paleoecologic zonation, the Lock Haven reef can be positioned within the biostratigraphic zonation (biozones) previously developed for these mid-Silurian units. F. M. Swartz (1970, pers. commun.) measured a section here many years ago when exposures were even less adequate than today. However, his section can be reinterpreted in terms of current understanding of this fossil reef (Fig. 3); in particular, he noted head-like colonies (reef core) at the level of the Velibeyrichia moodeyi biozone and branching colonies (deep reef flank) opposite the Whitfieldella marylandica biozone.

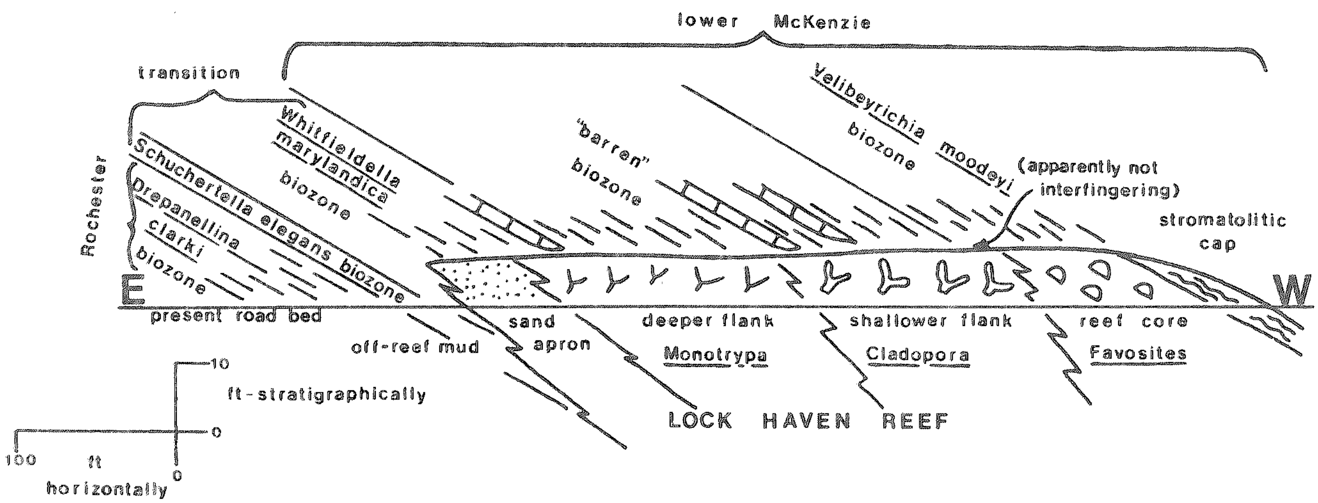

Figure 3. Coordination of ecozonation within, and biozonation adjacent to, the Lock Haven reef (adapted from Cuffey and Davidheiser, 1979, and F. M. Swartz, 1970, pers. commun.).

CONTINUE WALKING WESTWARD, FOR ANOTHER 350 FEET (106 METERS), BACK TO JUST BEYOND STOP 1, TO UPPER RIGHT (WESTERN) EDGE OF RUBBLY LIMESTONE EXPOSURE (STOP 7).

Stop 7: Stromatolitic Reef-Cap

Although now largely eroded away, a few blocks of massive, laminated, and stromatolitic limestone near the top of this exposure represent an algal reef cap developed after active reef growth had ceased, much as in certain Great Lakes Silurian reefs (Fig. 5J, N-P).

Light-gray to light-brown weathered, dark-gray fresh, and aphanitic, these limestones are homogeneous micrite (i.e., micstones) without shelly fossils. They are predominantly very low domed stromatolites (Collenia spp.), and form cruststones and bindstones. Such stromatolites suggest intertidal conditions, like the tidal flats covered by thin blue-green algal mats, in the Bahamas and Bonaire.

WALK WESTWARD AGAIN, 190 FEET (58 METERS) TO LIMESTONE EXPOSURE (STOP 8) AT ROAD LEVEL BELOW DOUBLE-BASED TELEPHONE POLE.

Stop 8: Rubble and "Oncolites" within Reef-Cap

Stratigraphically the topmost horizon of the Lock Haven reef, the bedrock here is a mixture of rubble breccia and bryozoan nodules or "oncolites" embedded in finer carbonate sediments. These materials seem most reasonably interpreted as reef-top or reef-flat rubble or cobble substrates (Fig. 5Q-S).

Varied in character, these are mostly dark gray fresh, weathering light gray to light brown, and aphanitic or finely crystalline; matrixes are diverse, including homogeneous micrite, pelleted micrite, and spar cement, while fossil contents range from negligible to abundant (shell fragments: brachiopods, gastropods, calcispheres, ostracods, and trilobites). Two variants, both rudstones, are especially notable among these reef-top limestones. One is composed of small round nodules, apparently algal oncolites, but actually rounded trepostome-bryozoan fragments (Monotrypa benjamini) (Fig. 5R-S), encrusted by vaguely cellular laminations (solenoporacean algae? chaetetid sponges? stromatoporoids?). Presumably, storm waters tossed broken bryozoan branches up from the deep flank, and those were further encrusted while being washed around upon the reef top. Another variant consists of subangular to rounded pebbles (up to 30 mm diameter) of favositid fragments, possibly a cemented reef-top rubble deposit (Fig. 5Q). The "oncolitic" and rubbly rock types suggest the cobble-strewn reef flats of Australian reefs like the Low Isles, intertidal and periodically quite turbulent.

STEP BACK NORTHWARD, ACROSS THE ROAD (BEWARE OF TRAFFIC!); FROM THE NORTH SIDE OF THE ROAD, LOOK OVER THE EDGE AND DOWN THE SLOPE (STOP 9).

Stop 9: Bryozoan "Oncolites" within Reef-Cap

Limestone float blocks composed of bryozoan "oncolites," again Monotrypa benjamini (Fig. 5R-S), weather out of the road fill below the pavement here. These apparently were originally part of the bedrock at the previous stop, but have since been displaced by road construction activities.

RETURN SOUTH ACROSS ROAD AGAIN (WATCH TRAFFIC!) TO ROADSIDE LIMESTONE EXPOSURE (STOP 8), AND CONTINUE ON WESTWARD FOR AN ADDITIONAL 215 FEET (66 METERS) ALONG THE SOUTHERN SIDE OF THE ROAD (STOP 10).

Stop 10: Favositid Satellite Bioherm

Approximately 110 feet (34 meters) horizontally south of, and 45 feet (14 meters) vertically above the roadside here is a small biohermal mound (Fig. 5T), interpreted as a satellite reef developed near the main Lock Haven reef mass. This mound consists of favositid coral heads embedded in a calcarenite matrix, but lacks any trepostome or cladoporid branch coquinas, and so resembles another small bioherm several miles farther east (Inners, 1984). Various other carbonate rock types, all non-reefal lithologies, can be seen in the bedrock exposures (apparently a small quarry) and float blocks in the mountainside woods surrounding the satellite reef.

WALK BACK EASTWARD, PAST ALL THE PREVIOUS STOPS, TO RETURN TO THE PARKING AREA: END OF FIELD TRIP NO. 1.

[see page 16]

Figure 4. Deeper-ecozone Lock Haven reef materials and modern analogues (scale of each photograph indicated as the actual width of the field of view shown). A-D, off-reef shale, Stop 5: A, platy shale and shaly limestone in vertical outcrop face (20 cm); B, modern analogue, shallow but far offshore carbonate mud resuspended in wake of shallow-draft research vessel out on interior of Great Bahama Bank (5-10 m); C, bedding surfaces of shale (7 cm); D, vertical peel-section of thin micstone bed (5 mm). E-H, calcarenite apron, Stop 4: E, massive grainstone weathering to sandy-textured surface (8 cm); F, modern analogue, barren sand ring around base of patch reef off St. Croix, grass-covered sand on left and base of reef on right (1.5 m); G, vaguely rippled bedding-plane surface (2 m); H, vertical peel-section (6 mm). I-R, trepostome flank, Stop 3: I, trepostome branch-coquina rudstone (9 cm); J, trepostome colony in growth position, embedded in calcarenitic matrix, thus forming bafflestone (11 cm); K, modern analogue, Acropora prolifera thickets off St. Croix (1 m); L, exposure of bedding plane containing in-place trepostome branching colonies, and C. Davidheiser on thickened mounded portion of bed (11 m); M, peel-section of trepostome branch-coquina rudstone, all Monotrypa benjamini fragments (11 mm); N-R, trepostome Monotrypa benjamini: N, half of large branching colony in upright growth position (48 cm); O, broken branch fragment showing characteristic finely tubular structure (13 mm); P, framework formed by colony branches in growth position (65 mm); Q, tangential peel-section (2.0 mm); R, longitudinal peel-section (2.0 mm).

[see page 17]

Figure 5. Shallow-ecozone Lock Haven reef materials and modern analogues (scale indicated as in Fig. 5). A-F, cladoporid flank, Stop 2: A, cladoporid branch-coquina rudstone (10 cm); B, modern analogue, Acropora cervicornis thicket off Grand Cayman (2 m); C, peel-section of cladoporid branch-coquina rudstone, all Cladopora seriata fragments (8 mm); D-F, tabulate Cladopora seriata: D, broken branch fragment showing typical coarsely tubular structure (14 mm); E, tangential peel-section (1.6 mm); F, longitudinal peel-section (2.3 mm). G-M, favositid core, Stop 1: G, favositid globstone (25 cm); H, modern analogue, Diploria heads on Bermuda (1 m); I, J, reef core as exposed in 1970 and 1983, respectively, with L. Zimmerman pointing to last remnant of stromatolitic reef cap (12 m, 3 m); K-M, tabulate Favosites niagarensis: K, head-like colony (50 cm); L, exterior surface approximating a tangential section (14 mm); M, longitudinal peel-section (7 mm). N-S, reef cap: N-P, stromatolite Collenia sp., Stop 7: N, very broad dome (40 cm); O, closely packed narrow columns (25 cm); P, vertical peel-section showing homogeneous micrite and subtle laminations (5 mm); Q, rubble-breccia rudstone, Stop 8 (25 cm); R-S, trepostome Monotrypa benjamini as rounded "oncolites" rather than branching colonies, Stop 9: R, "oncolite" rudstone (12 cm); S, tangential thin-section (3 mm). T, satellite bioherm west of main reef mass, Stop 10 (4 m).

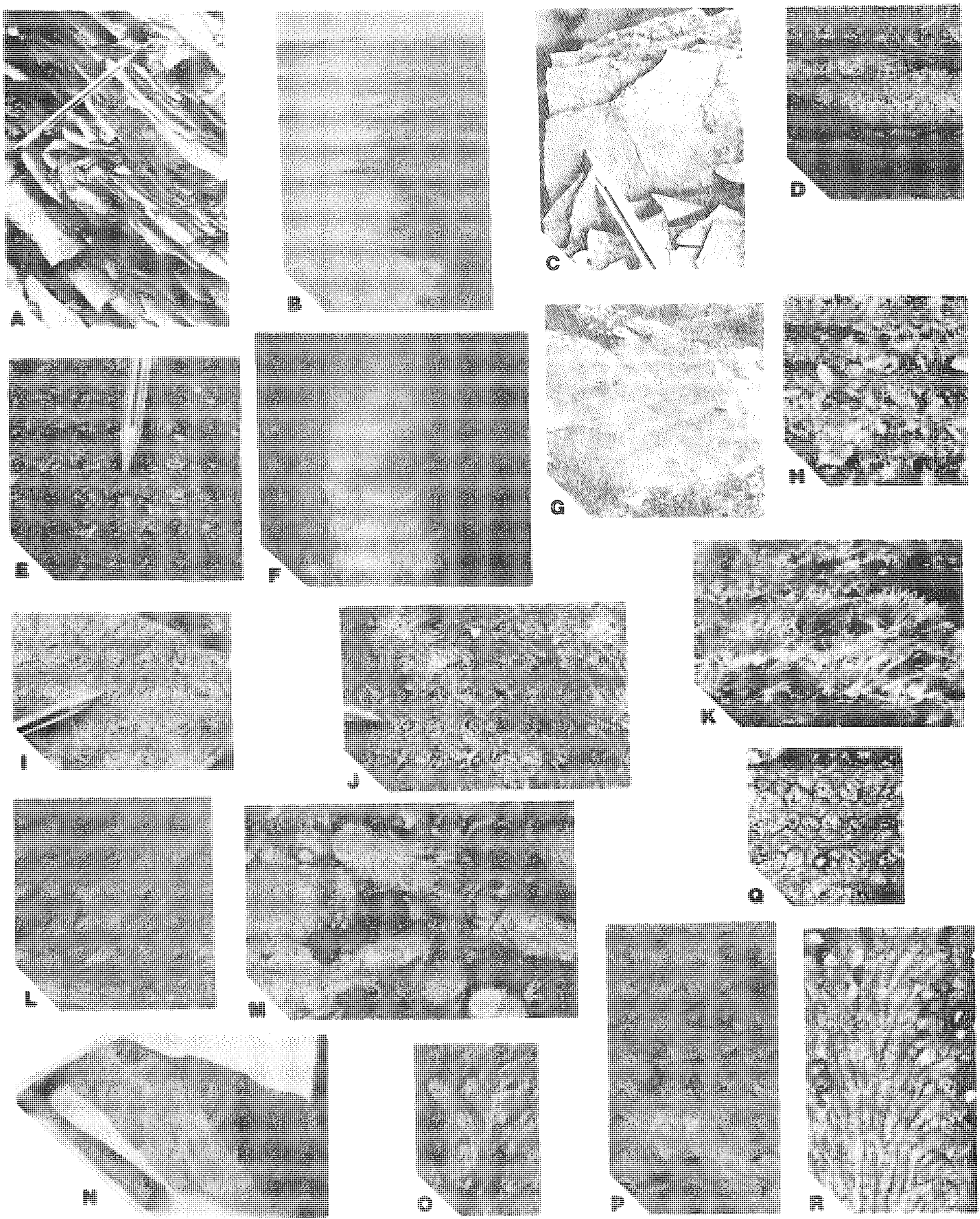

Figure 4.

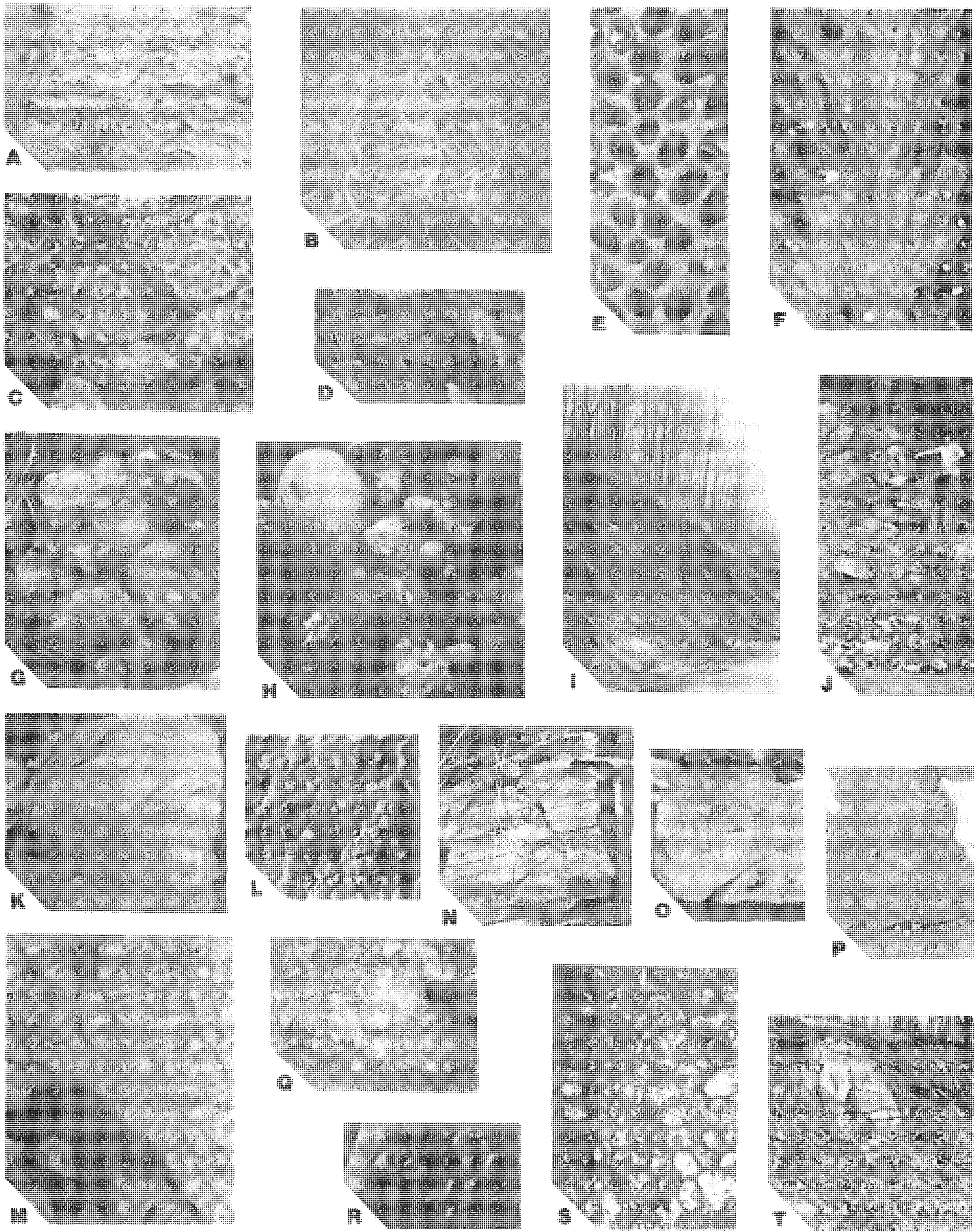

Figure 5.

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FIELD TRIP #2

CATSKILL SEDIMENTATION IN CENTRAL PENNSYLVANIA

Eugene G. Williams

(This report is a summary of the Ph.D. thesis by Victor Rahmanian, 1979).

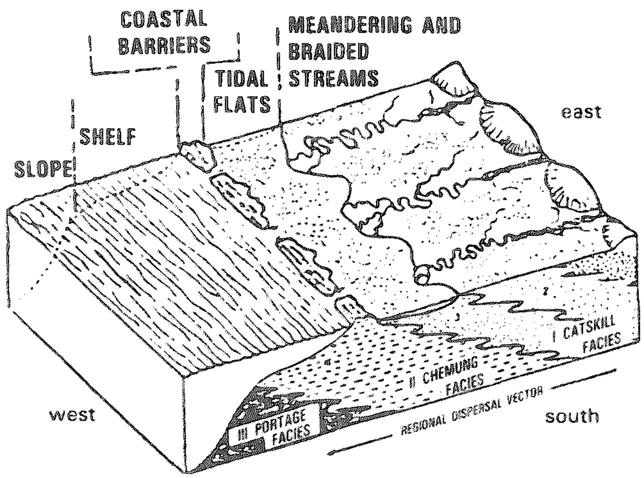
The Upper Devonian deposits of north-central and central Pennsylvania have been the subject of several geological studies in the past 20 years. Most of these studies contributed significantly to an improved understanding of the sedimentologic and stratigraphic aspects. However, the models of the depositional environment of these deposits have become quite controversial. The main point of disagreement arises from studies of Allen and Friend (1968) and Walker and Harms (1971), in the Susquehanna Valley area of south-central Pennsylvania (Fig. 1).

Allen and Friend (1968) concluded that the Catskill facies is not deltaic but instead was deposited in a vast coastal plain of alluviation located between the Acadian Mountains in the east and the Devonian seas in the west. This coastal plain stretched for at least 500 miles between New York and Virginia. It was characterized at its western margin by a network of barrier islands, tidal flats, and lagoonal environments, and at its eastern parts by an alluvial environment consisting of a network of meandering and braided streams (Fig. 1).

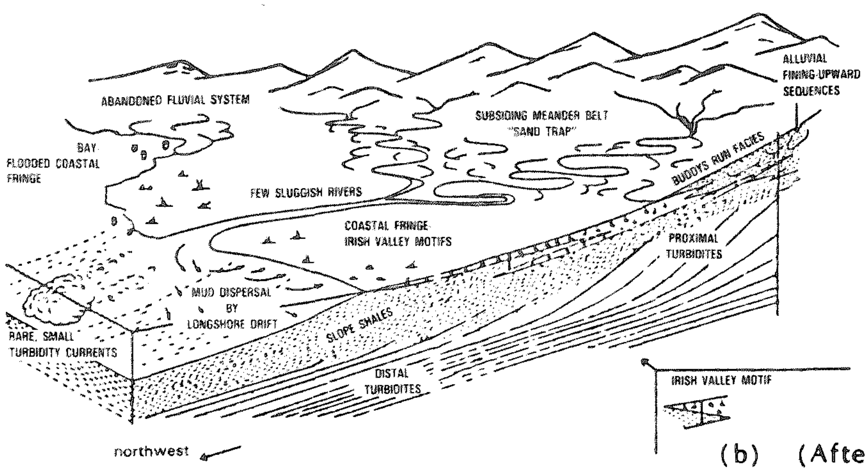
Walker and Harms (1971), in their study of the Upper Devonian of the same area, not only rejected Allen and Friend's interpretation but also questioned the deltaic nature of Catskill deposits in south-central Pennsylvania. They suggested that the Upper Devonian depositional system in this area was a quiet, prograding muddy shoreline, similar to muddy coastlines of southwestern Louisiana. They proposed that progradation of the shoreline was made possible by a supply of mud from a distant (unrecognized) delta by means of longshore currents. The sea supposedly had a low wave and low tidal range (Fig. 1). The principal criterion for this interpretation was the presence of several cycles, called "motifs" in the lowermost members of the Catskill Formation. These cycles were reported to consist of an alternation of marine-nonmarine sediments and to be devoid of prominent sand bodies. Winnowed sand bodies thicker than 50 cm were reported to be especially rare or absent in nearshore or shoreline positions of each cycle, suggesting a rather muddy shoreline which shifted back and forth through the course of sedimentation.

Glaeser (1974) studied the Upper Devonian sedimentary environments in northeasten Pennsylvania. He suggested a fluvially dominated deltaic model as the environment of deposition of Trimmers Rock and Catskill deposits (Fig. 1). In the proposed model by Glaeser (1974), prodelta, coastal-margin (delta-front and delta-plain) and alluvial-plain environments comprise the three major depositional environments of the indicated deltaic system. Using mostly subsurface information, he also constructed a three-dimensional time-stratigraphic correlation diagram and isopach maps which showed the regional relationship of the proposed deltaic components in northeastern Pennsylvania.

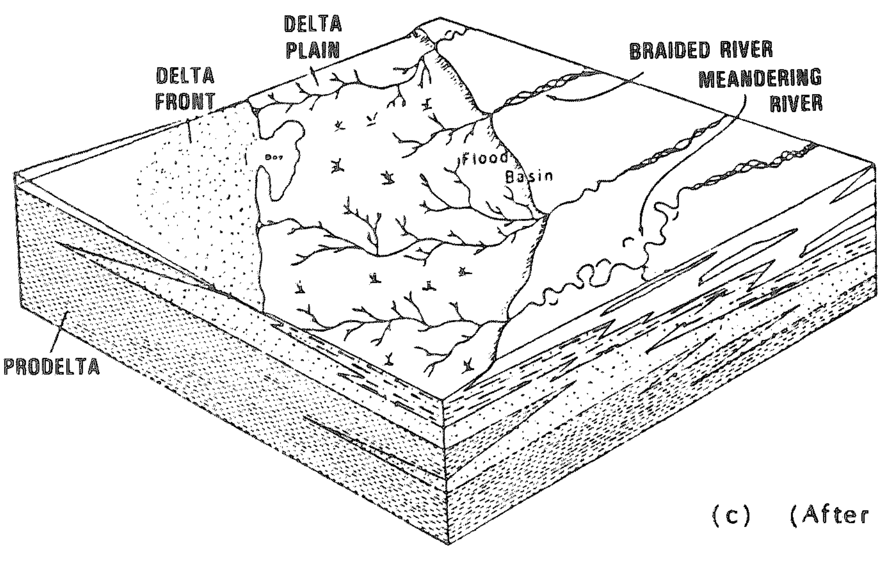
Humphreys and Friedman (1975) studied Catskill deposits in north-central Pennsylvania. They classified the Upper Devonian Catskill into three lithofacies: 1) gray marine sandstone and siltstones and shale, interpreted as of tidal origin; 2) coarse-grained gray-green, nonmarine sandstones, red siltstones and shale, inferred to have accumulated in and adjacent to meandering streams; and 3) coarse-grained, gray-green nonmarine sandstones of braided-stream origin.



(a) (After Allen and Friend, 1968).



(b) (After Walker and Harms, 1971).



(c) (After Glaeser, 1974).

Figure 1. Depositional models proposed for the Upper Devonian sediments of Pennsylvania.

In the study area, strata of the Catskill Formation are exposed along two almost parallel, northeast-trending outcrop belts, one along the northwest limb of the Broadtop syncline, and the other along the Allegheny Front in Centre, Blair, and Bedford Counties (Fig. 2). A northwestward paleoslope for the

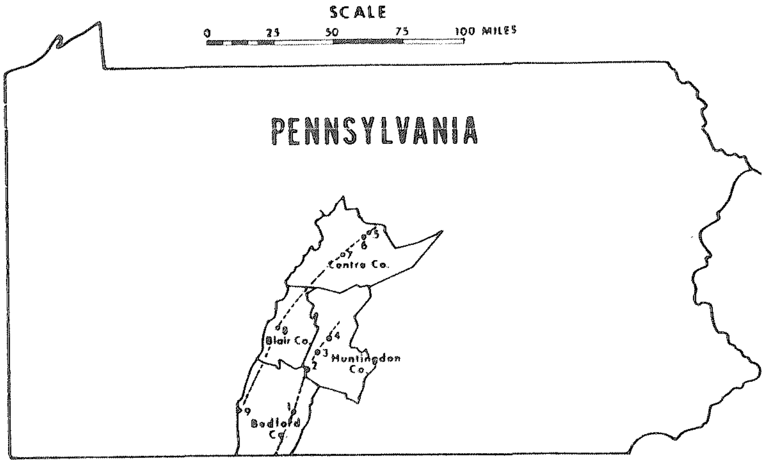


Figure 2. Index map of the study area. Numbers refer to the measured section (1-Everett, 2-Saxton, 3-Entriken, 4-Raystown Dam, 5-Milesburg, 6-Runville, 7-Port Matilda, 8-Horseshoe Curve, 9-New Baltimore). Dashed lines show the position of the Upper Devonian Catskill-Trimmers Rock formational boundary along the west flank of the Broadtop synclinorium (sections 1-4) and the Allegheny Front (sections 5-9) (from Rahmanian, 1979).

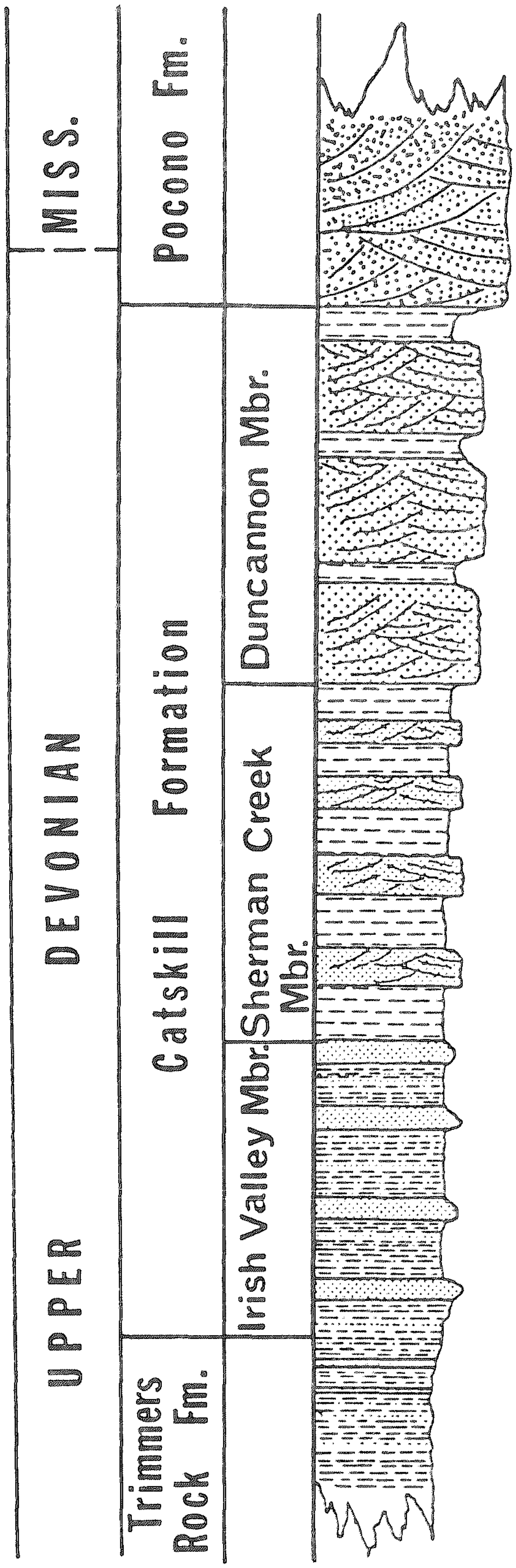
Upper Devonian depositional system has been documented by most of the studies on the Upper Devonian deposits in Pennsylvania and New York, and is confirmed by the paleocurrent measurements in the present study. Considering the northwestward paleoslope for these sediments, the northeast-trending outcrop belts of the study area provide a series of strike sections which give the opportunity to study the interrelationship of different depositional environments of the Upper Devonian rock in time and space. The average overall thickness of the Catskill Formation in this area is about 2000 feet, and the exposed thickness of this formation varies from 1200 to 1600 feet.

The Catskill Formation is divisible into three members in central Pennsylvania (Fig. 3). The basal Irish Valley Member consists of gray sandstone, siltstone, and shale, chocolate-brown siltstone and silty claystones, and red silty sandstone, siltstone, and claystone, all of which are arranged in several marine-nonmarine transitional cycles. The middle member (Sherman Creek Member) of this formation, where developed in the area, consists of a sequence of interbedded red siltstone and very fine grained sandstone, arranged in thin fining-upward cycles. The Duncannon Member consists of light-olive-gray and red sandstones, reddish-gray silty sandstone and red siltstone and silty claystone, arranged in well-developed thick fining-upward cycles.

Sedimentary units of Upper Devonian systems in the study area are the product of deposition in several environments identified as parts of a prograding coastline. Major facies comprising the Upper Devonian depositional system are shallow-shelf/littoral, chenier-plain/tidal-flat complex, tidal-flat/barrier complex, shelf-delta complex, and fining-upward alluvial complex.

The shallow-shelf/littoral facies is developed in the Trimmers Rock Formation, which underlies the Catskill Formation. These deposits consist of a sequence of thin- to medium-bedded and gray siltstone and olive-green to gray silty shale, interbedded with occasional thin layers of gray-green very fine grained sandstone layers. Crinoid columnals, pelecypods, brachiopod shells, and occasional bryozoan fragments are common in most of these beds. Except in two localities (Saxton and Port Matilda), the underlying Trimmers Rock sediments pass abruptly into nonmarine deposits without establishment of major nearshore sand bodies.

After the first establishment of nonmarine conditions, the sedimentary pattern of the Irish Valley Member of the Catskill Formation in the southern and central part of the study area is characterized by many (about 15-20) cycles consisting of



SUMMARY DESCRIPTION

Gray to buff, medium-grained crossbedded sandstone with lenses of white quartz-pebble conglomerate. Non-marine.

Asymmetric, fining-upward fluvial cycles, in which a basal, nonred, fine- to medium-grained, thick and crossbedded sandstone is overlain by very fine grained red sandstone, siltstone, and silty claystone (300 m).

Interbedded red claystone, siltstone, and very fine grained, crossbedded sandstone arranged in fining-upward cycles of fluvial origin (360-640 m).

Repeated cyclic alternation of olive-gray to reddish-gray marine sandstone, siltstone, and shale, and red nonmarine silt-stone, mudstone, and minor sandstone (180-300 m).

Olive-gray to gray marine silt-stone and silty shale with interbedded very fine grained sandstone in upper part.

Figure 3. Generalized stratigraphic section of the Upper Devonian in central Pennsylvania (from Rahmanian, 1979).

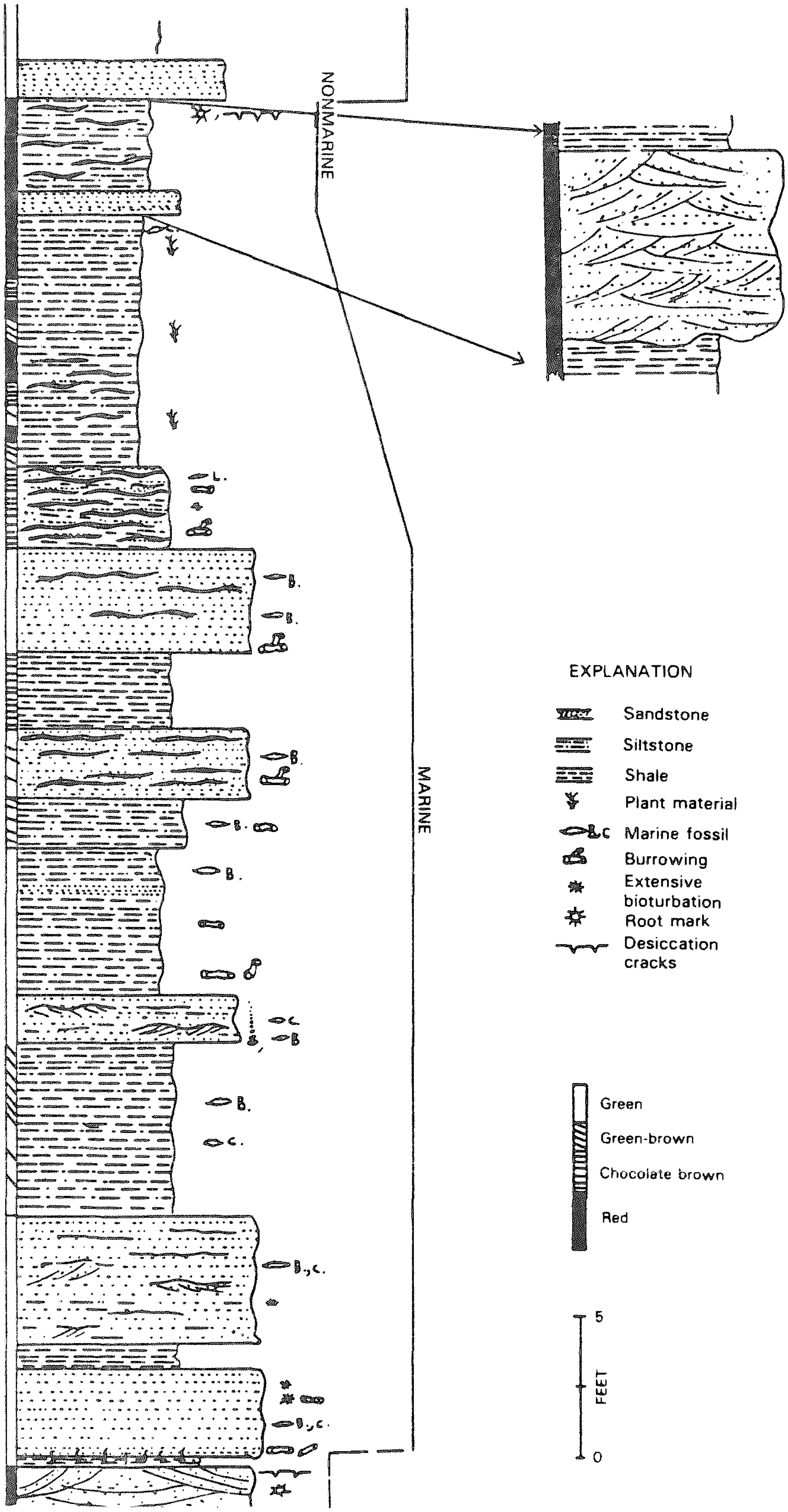


Figure 4. Generalized sequence of an Irish Valley motif from the Entriken section (from Rahmanian, 1979).

repeated alternations from marine sandstone and shale to nonmarine siltstone and silty sandstone which was produced by repeated lateral shifting of the shoreline. The thickness of each cycle varies from 5 to about 90 feet. These cycles begin with greenish-gray, fossiliferous, clean subparallel laminates overlain by bioturbated, fine-grained sandstone of variable thickness representing a marine transgression, and pass through a marine shoaling phase and an intertidal transitional phase, and finally grade into a nonmarine phase representing coastal-plain aggradation (Fig. 4). The marine shoaling phase of cycles commonly starts with gray-green to olive-green, fossiliferous shale and silty shale which grades upward to thin-bedded olive-green and chocolate-brown, fossiliferous and bioturbated shaly siltstone occasionally interlayered with thin layers of gray-green, very fine grained, fossiliferous, micro-cross-ripple-laminated sandstone. The shoreline of this marine shoaling phase is represented by usually thin (2-5 feet), olive-green, fine-grained, moderately sorted, subparallel to flaser and lenticular laminated, fossiliferous quartzitic sandstone. The transitional part of each cycle usually consists of interlayers of green, chocolate-brown, and red siltstone, shaly siltstone, and thin (1/2-1 foot) fine-grained, clean, well-sorted quartzitic sandstone.

Extensive bioturbation, diagnostic internal sedimentary structures such as herringbone cross-stratification, lenticular and flaser bedding and presence of composite rock types in the shallow marine-transitional part of most cycles attest to tidal origins of these deposits. Tidal sedimentation in some of the sections is further demonstrated by good development of relatively thick (5-20 feet), gray-green to chocolate-brown, medium to coarse and pebbly, fossiliferous cross-stratified quartzitic sandstone and conglomerate interpreted as tidal channels, and subtidal and intertidal sand bars.

The nonmarine part of cycles is dominantly red siltstone and shale which are characterized by the presence of rootlets and mudcracks. Fining-upward alluvial cycles of a few feet in thickness may be present on top of some cycles.

Upward-fining cyclicity of fluvial origin is the common characteristic of the other two members of the Catskill Formation (the Sherman Creek and Duncannon Members). An ideal cycle consists of a basal brownish-gray to red, fine- to very fine grained, micaceous, crossbedded sandstone with lenses of carbonate nodules and shale chips, and occasional plant fragments at its base. This sandstone occupies a channel or irregular erosional surface cut into the underlying cycle. This sand body grades upward to red to reddish-gray, very fine grained silty sandstone, red siltstone, and silty shale which represents the levee-overbank portion of a meandering-channel facies.

Facies assemblages and distribution of the Catskill Formation vary both towards the south and north laterally along depositional strike from the above description for the central parts of the area (Figs. 5 and 6). To the south, at the Saxton area, the first nonmarine deposition, marking the base of the Irish Valley Member, was established higher in the section, and consequently shallow marine sedimentation went on for a longer period of time relative to the adjacent area. Apparently this was in response to a reduction of the rate of sediment supply, which more or less prevented active coastal progradation and consequently resulted in more intensive marine reworking and a better development of shallow marine and nearshore sand bodies. In this area the sediments of the uppermost Trimmers Rock Formation and the Irish Valley Member are represented as tidal-flat and barrier-bay facies assemblages characterized by better developed and thicker marine sandbars and shoreface-foreshore sequences with associated tidal-channel sandstone bodies.

To the north (at Centre County) the well-developed cyclic sediments of the Irish Valley Member grade into a complex assemblage of tidal-influenced deltaic facies, comprising slope, prodelta, delta-front, and lower and upper delta plain facies (Figs. 5 and 6).

In summary, the available data on Upper Devonian deposits of the study area suggest that the depositional system of Late Devonian time in the south-central parts of Pennsylvania was a complex prograding delta-interdeltaic system (Fig. 7). The shoreline was fed by a tidally influenced delta at the northern part of the study area (at Centre County). Farther to the south and north of this depocenter occur well-developed cycles of marine-nonmarine origin in the Irish Valley Member, which point to development of a prograding, tidally influenced, muddy shoreline, consisting of a tidal-flat/chenier-plain complex, marginal to the delta. Available evidence suggests that sediments were supplied to these environments from the adjacent deltaic lobe by longshore currents and tidal currents rather than by local rivers crossing the coastal plain.

This shoreline shifted laterally several times in response to changes in position of the adjacent deltaic lobe. Further to the south, at the Saxton area, cyclic sediments are poorly developed. In this area, marine processes became dominant, as the rate of sediment supply through longshore currents from the north was reduced and a tidal-flat/barrier-bay complex formed the shoreline of this area (Fig. 7).

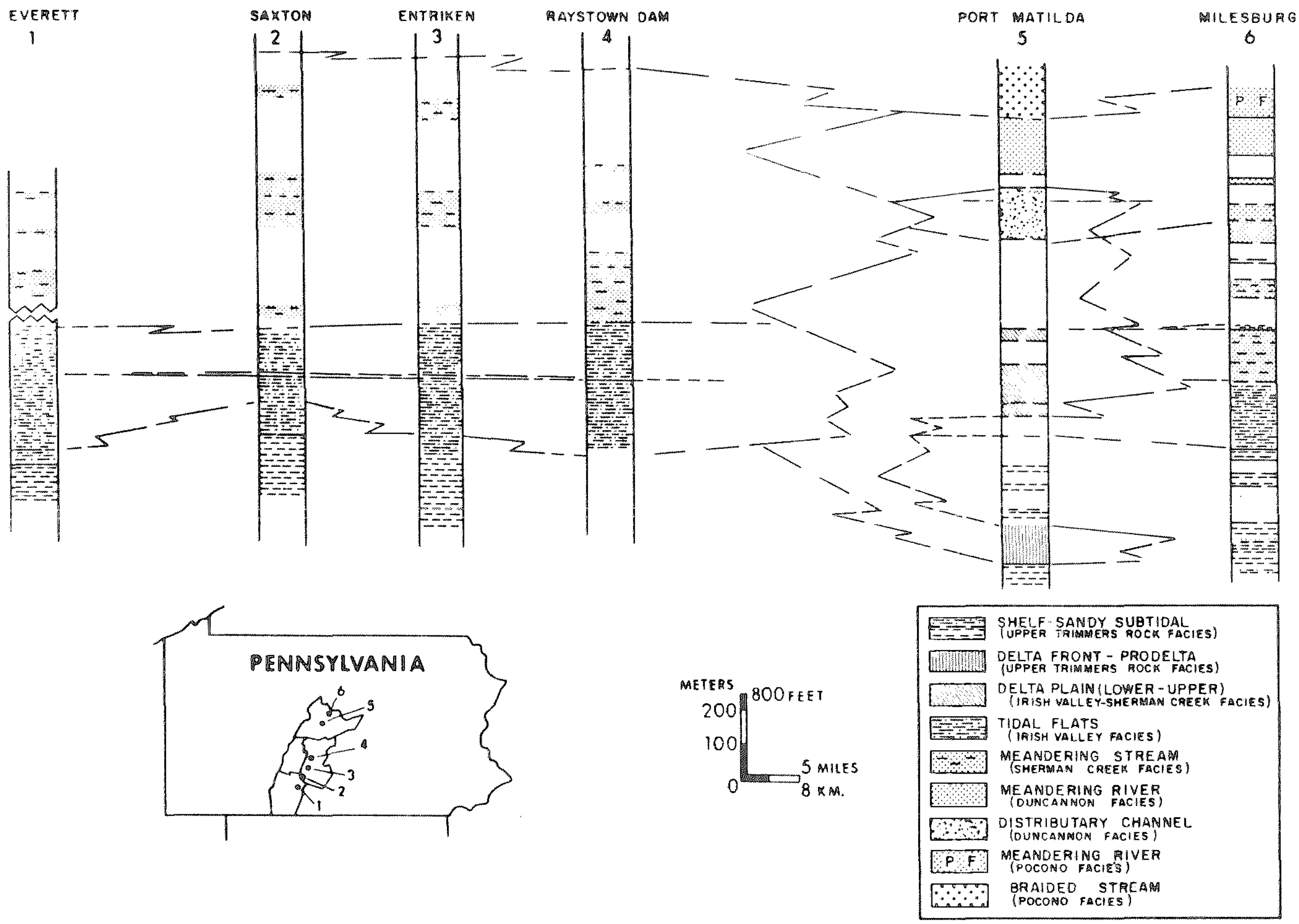


Figure 5. Correlation diagram of the major lithofacies in the Upper Trimmers Rock and Catskill Formations of central and south-central Pennsylvania and the inferred paleoenvironments (from Rahmanian, 1979).

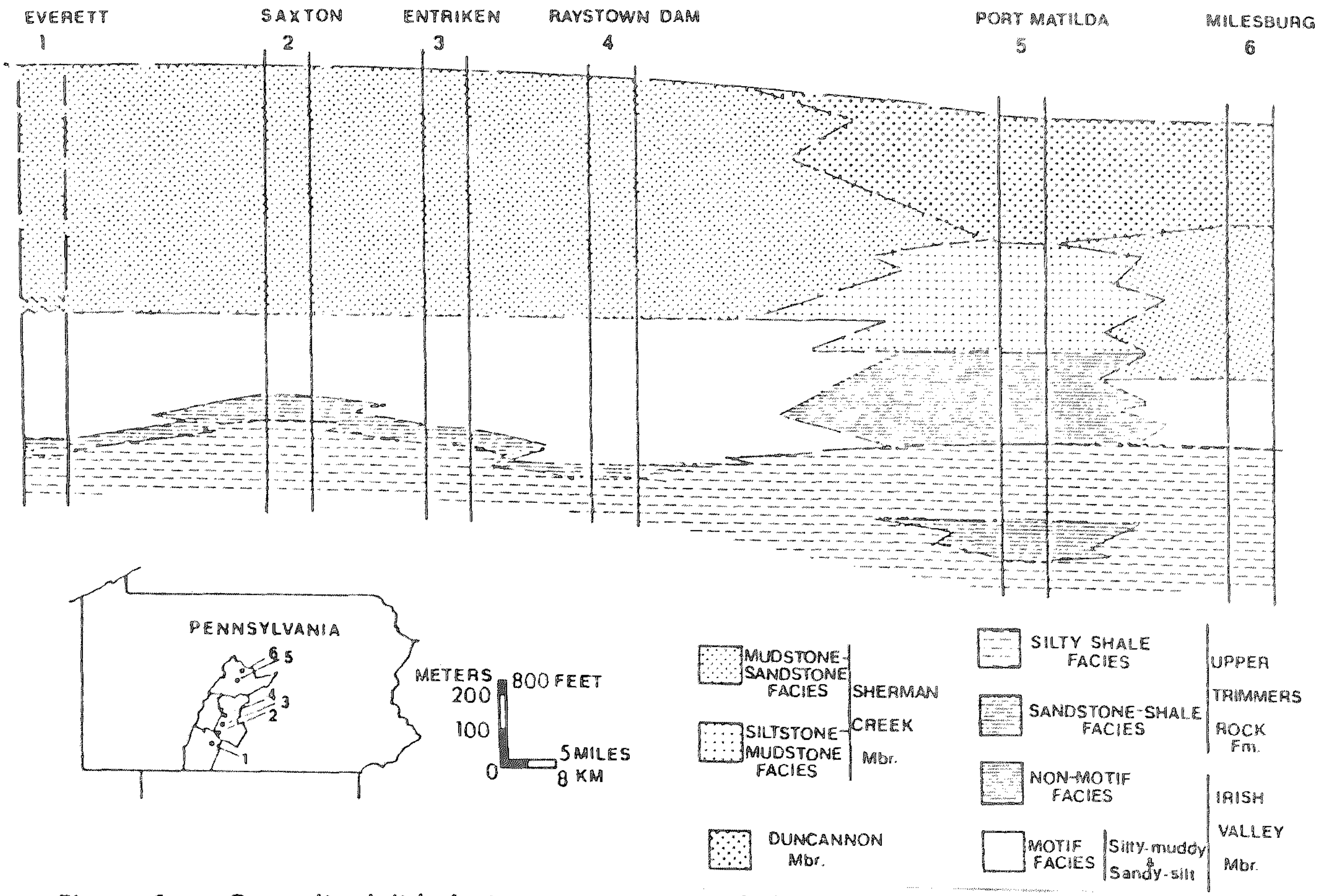


Figure 6. Generalized lithofacies cross section of the Upper Devonian rocks in central and south-central Pennsylvania (from Rahmanian, 1979).

At a given locality, for example along the Broadtop outcrop belt, the vertical succession of sedimentary facies indicates that the area, during the entire time of Irish Valley sedimentation, remained a part of an intradeltaic region, characterized by a coastal area largely composed of broad and extensive tidal flats. The lack of major sandstone units of fluvial or deltaic origin in the succession of these marine-nonmarine deposits indicates that the shoreline was not subjected to intermittent deltaic or intradeltaic sedimentation. In other words, neither the rivers from adjacent active sediment input systems nor a major river, heading directly from the southeastern source area, crossed this shoreline during Irish Valley time. This evidence leads to the important conclusion that the paleogeographic position of the major facies of the Catskill Formation (i.e., Irish Valley, Sherman Creek, and Duncannon) remained the same throughout the complete progradation history of the Upper Devonian across central Pennsylvania.

More specifically, the evidence suggests that the main river system in the north-central part of the area maintained a fixed longitudinal course without diversion or extensive lateral migration as it crossed the Upper Devonian basin across central Pennsylvania. The inferred longevity of the river system or confinement of its course suggests that structural elements possibly influenced its course during its westward journey across the Upper Devonian sedimentary basin. The river system was probably confined to a structurally low area which controlled its course and limited its lateral migration. A likely structural element which could have exerted such control on the course of this river system is the NW-SE-trending basement-

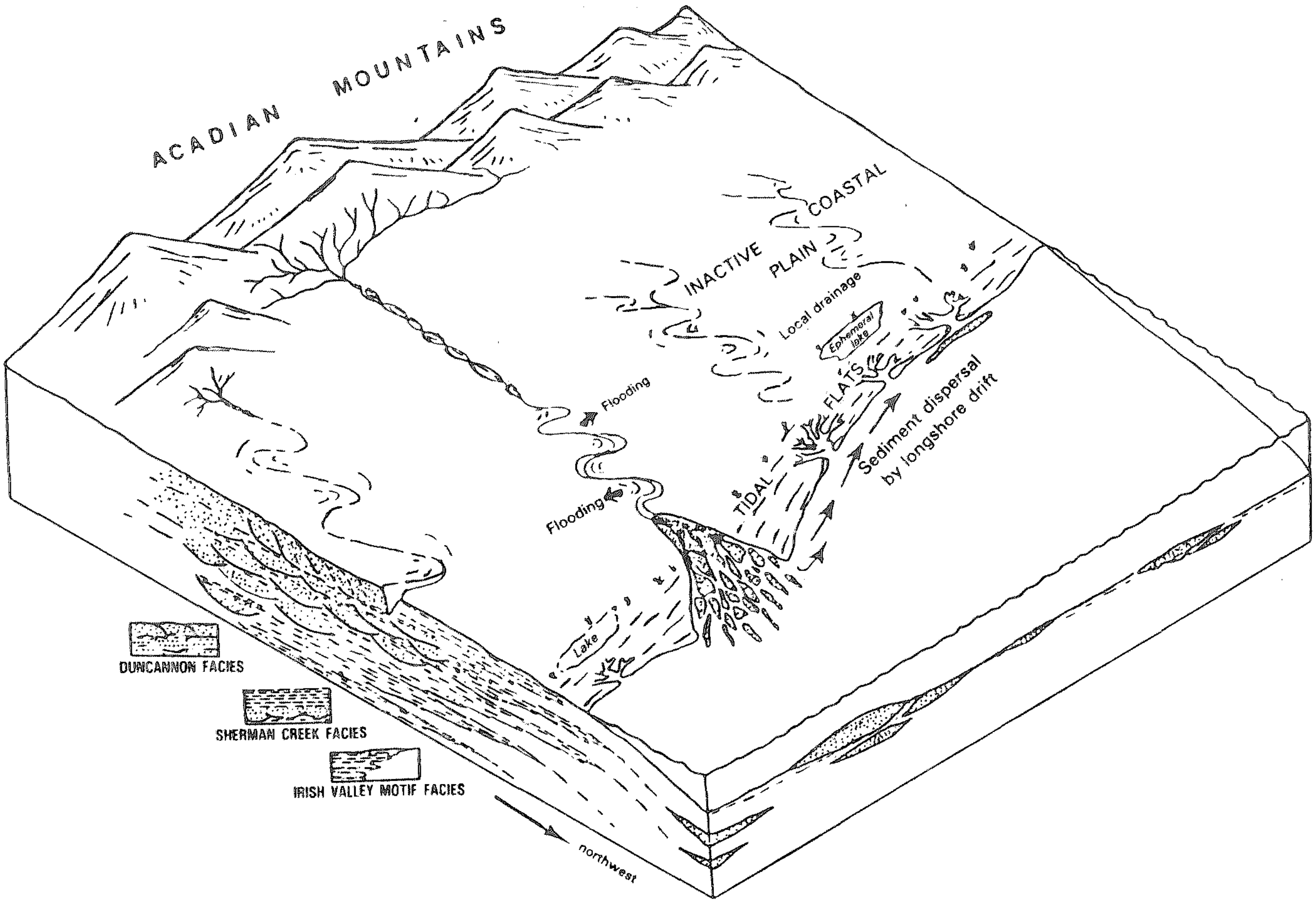


Figure 7. Sedimentation model of the Upper Devonian in central Pennsylvania (from Rahmanian, 1979).

controlled fault system mapped as the Tyrone-Mount Union lineament (Canich, 1977). This fault system extends almost perpendicular to the inferred depositional strike of the Devonian basin and separates the inferred sediment input area at Centre County in the north from its adjacent inactive coastal plain and marine embayments to the south.

The tidally dominated delta and interdelta shoreline model of Catskill sedimentation in central Pennsylvania was tested by two independent lines of evidence, namely paleocurrent analysis and petrography of sandstones.

Figure 8 illustrates the paleocurrent directions from crossbeds of the entire Irish Valley, Sherman Creek, and Duncannon Members at various localities. Figure 8 also illustrates the outline of the Devonian outcrop belts in the study area. Paleocurrent azimuths measured from the Sherman Creek and Duncannon Members are characteristically unidirectional and indicate progradation of nonmarine sequences in WNW to dominantly NW directions with respect to a NE-SW shoreline and a southeasterly source terrane. As mentioned in the preceding section, these sediment transport directions are in general agreement with regional paleodirectional measurements conducted by various workers on the Upper Devonian deposits of New York, Pennsylvania, and Maryland.

The paleocurrent azimuths from the crossbeds of the Irish Valley Member sandstones clearly indicate a NW-SE bipolar current direction (Fig. 8). A bipolar

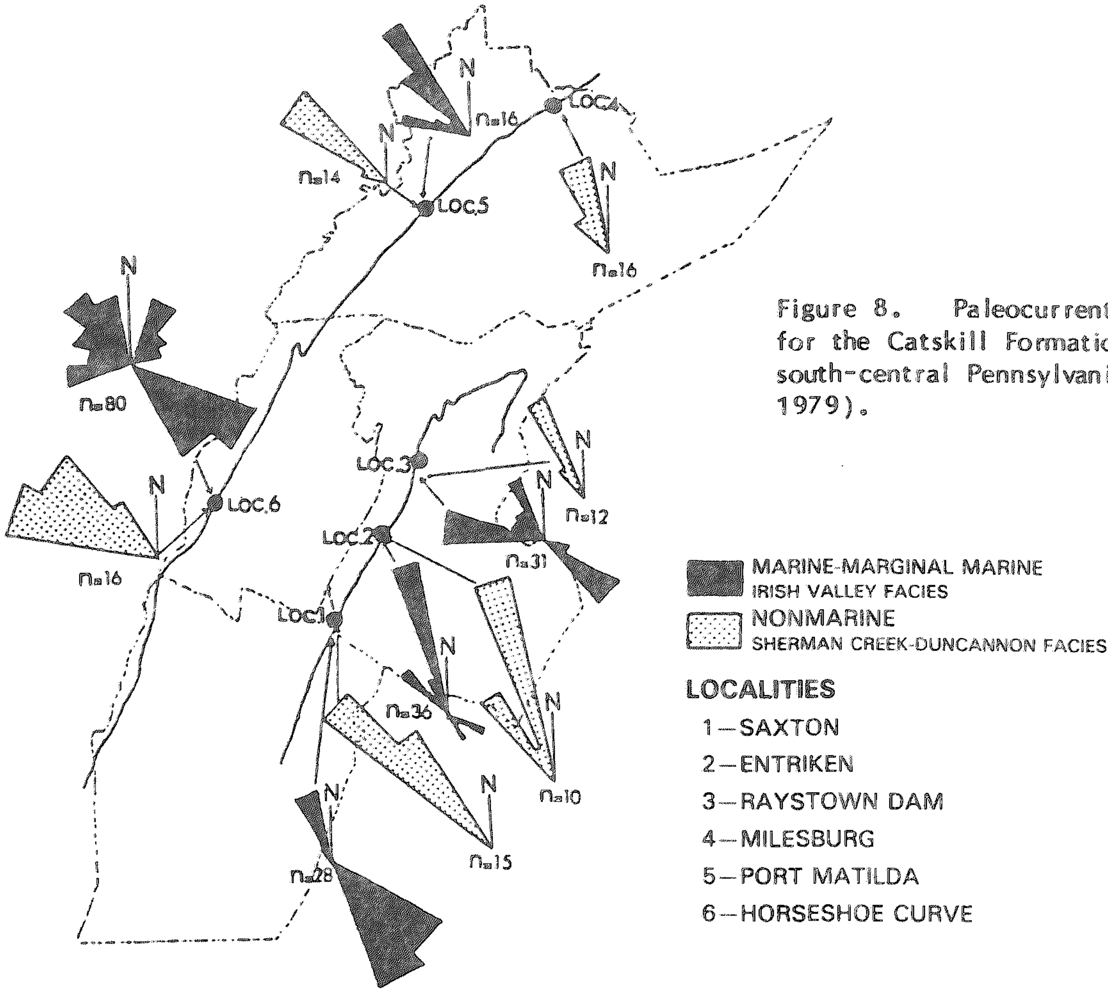


Figure 8. Paleocurrent directional data for the Catskill Formation in central and south-central Pennsylvania (from Rahmanian, 1979).

flow pattern inferred from crossbedding in this member is most satisfactorily accounted for by the ebb- and flood-current reversal typical of tidal environments. These paleodirectional data provide additional support for a tidal origin of the Irish Valley motif sequences. The relationship of the crossbed orientation in the Irish Valley sandstones (Fig. 8) with the inferred NE-SW shoreline trend indicates that the northwesterly and southeasterly oriented crossbeds are ebb- and flood-generated, respectively. Paleocurrent analysis of the Irish Valley sandstones (Fig. 8) suggests that in the Raystown Dam and Entriken areas the ebb currents have been the dominant constructional process in addition to a subordinate flood-current component, while the Saxton area appears to represent a flood-dominated shoreline with subordinate ebb-current components. These relationships are suggested by both the composite paleodirectional measurements of all the Irish Valley motif sandstones, and individual sandstone bodies which show a distinct unimodal, ebb- or flood-generated flow pattern. An example of such individual units is the thick conglomeratic sandstone unit at the Entriken section, which has distinct northwesterly oriented large-scale solitary sets generated by tidal-flood processes. Paleodirectional data at the Horseshoe Curve locality, while characterized by well-defined ebb and flood paleocurrent components, also show a distinct northeasterly oriented paleocurrent direction. This is interpreted to have been generated by tidal processes combined with a northeasterly long-shore movement of water. Similar influences of longshore movement of water have been inferred from the results of the paleodirectional measurements from the subtidal sandstone units of the uppermost

Trimmers Rock Formation at Saxton. Paleodirectional data of the Irish Valley Member at Port Matilda are exclusively from the sandstone units interpreted as distributary-mouth bars and show, predictably, a northwestward paleocurrent direction.

The petrographic analysis was based on 58 samples taken from sandstone units of the Trimmers Rock and Catskill Formations. These sandstone units are interpreted, based on field evidence, to represent sedimentation within five depositional environments, including shelf, subtidal/tidal-channel, beach, lower tidal flat, and fluvial. The mean values and standard deviations of the properties measured are given in Table 1. Examination of this table will show that these sandstones can be separated into three groups, namely marine shelf, shoreline (beach, tidal channel,

Table 1. Group means and standard deviations for each variable for the several paleoenvironments identified in the Catskill and Trimmers Rock Formations (size in phi; composition in percentage) (from Rahmanian, 1979).

Variables	Marine shelf		Beach		Subtidal-tidal channel		Tidal flat		Fluvial		Grand
	$\bar{x}$	S.D.	$\bar{x}$	S.D.	$\bar{x}$	S.D.	$\bar{x}$	S.D.	$\bar{x}$	S.D.	mean
1. Mean size a-axis	3.453	0.601	2.688	0.534	2.135	0.739	2.756	0.179	3.352	0.566	2.957
2. Standard deviation a-axis	0.400	0.070	0.386	0.087	0.748	0.367	0.369	0.042	0.420	0.070	0.466
3. Maximum size a-axis	2.570	0.674	1.954	0.612	0.173	1.373	1.900	0.268	2.518	0.499	1.911
4. Mean size b-axis	4.067	0.553	3.226	0.542	2.657	0.762	3.308	0.182	3.965	0.580	3.533
5. Standard deviation b-axis	0.432	0.083	0.388	0.087	0.761	0.348	0.396	0.041	0.424	0.087	0.480
6. Maximum size b-axis	3.187	0.582	2.364	0.615	0.689	1.345	2.410	0.348	3.159	0.564	2.470
7. Axial ratio (b/a)	0.676	0.031	0.709	0.034	0.715	0.030	0.699	0.014	0.678	0.03	0.692
8. Monocrystalline quartz	47.389	2.863	59.741	4.106	61.448	8.907	57.875	6.810	45.740	4.55	52.907
9. Polycrystalline quartz	5.331	2.581	6.515	3.195	13.105	8.346	4.207	1.486	4.683	2.110	6.633
10. Feldspar	2.556	0.999	3.221	1.520	2.240	1.421	3.749	2.364	3.44	2.011	3.039
11. Mica	4.248	3.122	1.036	0.874	1.061	0.988	0.498	0.471	3.713	1.901	2.459
12. Metamorphic rock fragment	18.083	3.940	5.702	2.705	6.847	5.128	12.625	7.181	17.964	4.289	13.241
13. Sedimentary rock fragment	0.194	0.265	0.740	1.391	0.333	0.536	1.581	1.178	0.760	0.814	0.672
14. Volcanic rock fragment	0.000	0.000	0.111	0.236	0.393	0.679	0.000	0.000	0.000	0.000	0.092
15. Detrital matrix	11.027	5.977	1.703	1.305	2.031	2.289	3.249	3.951	8.556	5.225	6.034
16. Authigenic matrix	1.278	1.154	1.038	1.060	0.515	0.970	2.375	1.148	0.648	0.621	1.052
17. Silica cement	8.557	3.411	12.260	5.195	7.056	1.873	9.709	3.098	6.926	2.122	8.500
18. Carbonate cement	0.027	0.095	7.370	5.630	4.030	5.957	3.710	4.941	0.000	0.000	2.425
19. Iron cement	0.833	1.789	0.037	0.110	0.787	1.752	0.209	0.590	7.315	3.912	2.626
20. Heavy minerals	0.471	0.480	0.517	0.912	0.151	0.229	0.206	0.171	0.258	0.293	0.315
21. P-M ratio (poly/mono)	0.112	0.053	0.108	0.047	0.224	0.154	0.076	0.034	0.102	0.041	0.124

tidal flat) and fluvial. The fluvial and marine shelf sandstones are quite similar in grain size (very fine grained), whereas the shoreline sandstones are all fine grained, a condition which suggests that the latter were not derived from the former. Because most of the fluvial samples were taken from the Sherman Creek, whose paleogeographic position was immediately upslope from the shoreline sediments of the Irish Valley, we conclude that these latter were delivered by longshore currents carrying sediments derived from coarser grained sediments of depocenters located near Port Matilda to the north or somewhere in Maryland to the south.

As expected, total quartz is higher and detrital matrix, mica, and metamorphic rock fragments lower in the shoreline sandstones when compared to the shelf and fluvial ones in which these components occur in about equal amounts. The higher energy and longer residence time in the nearshore environments probably accounts for the differences. Other important differences are found in carbonate and iron cements; carbonate cement, in the form of calcite, is much more abundant in the nearshore sandstones, very low in the shelf, and absent in the fluvial ones, relations which are probably a function of pH and permeability. Higher pH and permeability in the nearshore sandstone would favor the formation of calcite during diagenesis as would access of calcite-saturated seawater and opportunity for degassing of CO_2. In contrast, the lower pH in fluvial environments would likely favor calcite solution. In the shelf facies, higher matrix means lower permeability and therefore little space for cementation regardless of the availability of carbonate. The high iron content, in the form of hematite, in the fluvial sandstone, is explained by the fact that iron is generally insoluble in oxidizing environments.

The sandstones may be further separated by diagrams relating composition and texture. Figure 9 shows that the shoreline sandstones, in addition to being on average coarser grained, are also higher in quartz for a given grain size when compared to the fluvial and shelf sandstones. Figure 10 shows that the proportion of polycrystalline quartz relative to monocrystalline quartz is lowest, for a given grain size, in the shoreline sands, a feature to be expected in sediments where residence time is the longest so that differential abrasion and selective sorting are most pronounced.

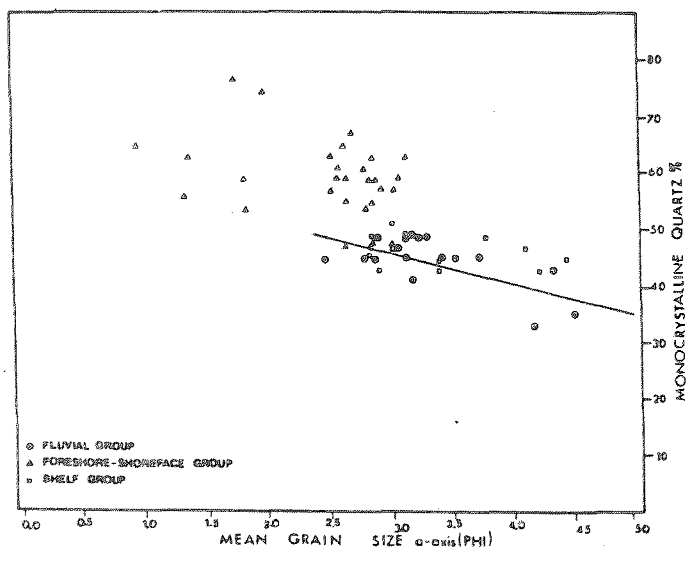


Figure 9. Plot of mean grain size of sandstones from the Catskill and Trimmers Rock Formations against monocrystalline quartz percentage (from Rahmanian, 1979).

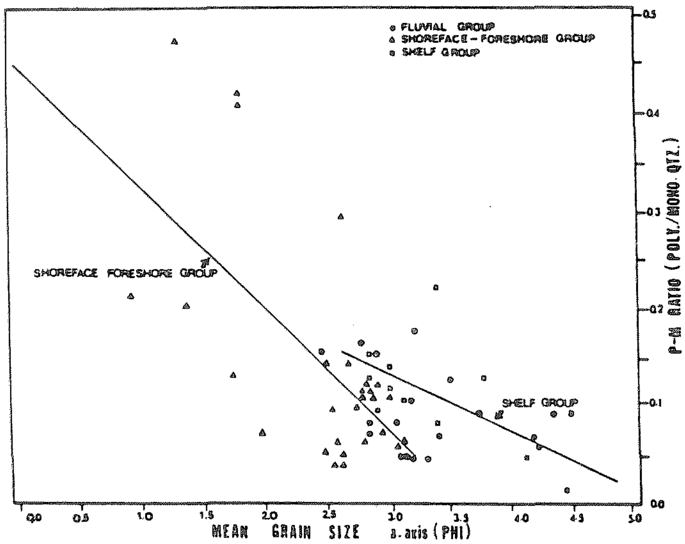


Figure 10. Plot of mean grain size of sandstones against P-M ratio (polycrystalline quartz/monocrystalline quartz) (from Rahmanian, 1979).

Our concept of Catskill sedimentation in central Pennsylvania is summarized in the idealized diagram presented in Figure 7. Reference to Figure 1 will show that the model incorporates as special cases the three suggested models which had been developed for specific areas in other parts of Pennsylvania. We therefore conclude that the differences in interpretation of Catskill sedimentation are largely related accidents of outcrop--each model valid for the area investigated but not for the whole region. The excellent exposures on Route 322 west of Port Matilda have allowed us to recognize facies in the upper Trimmers Rock and Catskill that had not been observed in this area.

Recent work by Smith and Rose (1985) has applied this model to the exploration for uranium in Pennsylvania, the results of which are summarized in this volume. They recognize another depocenter in northeastern Pennsylvania with characteristics similar to those found at Port Matilda, thus supporting the idea of Willard (1939), namely that there were multiple depocenters. Willard, based on biofacies analysis in the upper Trimmers Rock, concluded that there were three delta lobes, one of which (the Snyder Lobe), would, when projected westward, correspond to the inferred Port Matilda delta sequence.

SEDIMENTARY PROCESSES ON THE BASIN MARGIN OF THE LATE DEVONIAN CATSKILL SEA -- STORM- OR TIDE-DOMINATED?

by

Rudy Slingerland

Introduction

A central thesis of Rahmanian (1979) and Williams (this volume) is that the Upper Devonian Catskill shoreline in central Pennsylvania consisted of discrete deltas, each building onto a basin margin influenced by tide- and storm-driven flows. The assertion is made primarily from field observations given in the accompanying article and field guide. Proving the thesis is especially difficult in this case because there seems to be no modern counterpart to the Catskill depositional system. The facies sequences do not match modern river-dominated deltas, and so the specific coastal geomorphology is debated--were deltaic depocenters present, forming an irregular, embayed coastline or was the coastline muddy and relatively straight (see Williams, this volume, for a review)? Also, the submarine topography of epicontinental seas, and the Catskill Sea in particular, is poorly known. I have adopted Woodrow and Isley's (1983) term, basin margin, for the shallow edges of the Catskill Sea to emphasize its potential differences from modern continental shelves. Again for lack of modern counterparts, the occurrence of epicontinental tides in general, and Catskill tides in particular, is debated. Dennison (1985) and Woodrow and Isley (1983) ruled out tidal flows in the Catskill Sea on intuitive grounds because it was partly enclosed and comparatively shallow. Finally, on a basin margin as extensive in space and persistent in time as the Catskill, and which experienced numerous transgressions and regressions, it is entirely possible that both storm- and tide-driven flows were dominant at different times and places, depending upon the basin margin geometry. Certainly, this idea is suggested by the widely different process interpretations of Upper Devonian sedimentary facies in New York and Pennsylvania (Table 1).

Among the many questions raised above, two are both important and addressable now: 1) are significant tidal ranges even theoretically possible in the Catskill Sea, given our best estimates of basin shape, bathymetry, latitude, open ocean tidal ranges, and paleo-geophysical constants; and 2) how would those ranges vary at the coast as a function of basin margin widths, lengths, and depths? The purpose of this paper is to attempt to answer these questions by presenting the results of a numerical model simulating two-dimensional, shallow-water, long-wave propagation.

Methodology

To address the first question, a numerical hydrodynamic model simulating an M_2 co-oscillating tide was modified from Hess and White (1974) for the Catskill Sea (Slingerland, 1984, 1985, 1986). The model consists of the vertically integrated Navier-Stokes and continuity equations in which bed stresses are represented by the Chezy relationship where Manning's n equals 0.04 (gravel roughness). Basin planform was adopted from Heckel and Witzke (1979). Basin bathymetry was estimated from the facies maps of Heckel and Witzke (1979) using the basin margin-clinoform-basin floor scheme of Woodrow and Isley (1983). Forty-nine numerical experiments were conducted using a range of boundary and initial conditions. Solutions consisted of a vertically averaged flow velocity in the horizontal plane and water surface elevation

Table 1. Previous process interpretations for basin margin deposits in the Late Devonian Catskill Sea

Rock unit and investigator	Location	Specific bed type	Environment or mode of emplacement
West Falls Gp. (Woodrow and Isley, 1983)	Wellsburg, N.Y.	Hummocky cross-stratified fine sandstones	Wind-driven currents
Lock Haven Fm. (Woodrow and Isley, 1983)	Towanda, Pa.	Fossiliferous channelized sandstones	Flood discharges across delta platform from adjacent distributary
New Milford Fm. (Krajewski and Williams, 1971)	Susquehanna County, Pa.	Plane-parallel lam., flat-based sandstone pods in mudstone	Wave-swash on beaches or tidal flows
Irish Valley Mbr. of Catskill Fm. (Allen and Friend, 1968)	Newport and Girtys Notch, Pa.	Fining-upward shelly sandstones	Lateral accretion in tidal channels
Irish Valley Mbr. of Catskill Fm. (Walker, 1971)	do.	do.	Distributary fill or channel-mouth bars
Sonyea Gp. (Sutton and others, 1970)	Ithaca, N.Y.	Crossbedded fine sandstones with coquinite bases and wave-rippled tops	Storm, tidal, or wave-produced currents on a delta platform
Sonyea Gp. (Goldring and Bridges, 1973; Duke, 1985)	N.Y.	Sublittoral sand sheets with hummocky cross-strata	Hurricane-driven currents
Bioclastic carbonate units in Lock Haven Fm. and West Falls Gp. (Woodrow and others, 1981)	NE Pa.	Bidirectional trough and wedge crossbedded skeletal-fragment lime grainstone with lateral accretion bedding	Ebb-dominated tidal environment
Ashcraft unit (see above) (Bridge and Droser, 1985)	do.	do.	Laterally migrating sand bars adjacent to tidal channels
Sonyea Gp. (Goldring and Langenstrassen, 1979)	N.Y.	Sandy shell layers truncating bioturbated mudstones	Storm layers deposited rapidly under foul weather conditions
First Bradford Fm. (Murin and Donahue, 1984)	SW Pa.	Very fine to fine sandstone pods up to 10 m thick and oriented NE-SW	Inner to mid-shelf bar facies accumulated by current reworking
Upper Devonian hydrocarbon reservoir sandstones (Donaldson and others, 1984)	NW W. Va.	Herringbone crossbedded channel sandstones	Tidal flows

(with respect to mean water level) as functions of location within the basin, phase of the tide, and open ocean tidal range at the entrance to the basin. The results of one experiment, based on the present best estimates of basin geometry and Devonian open ocean tides, are presented in Figure 1.

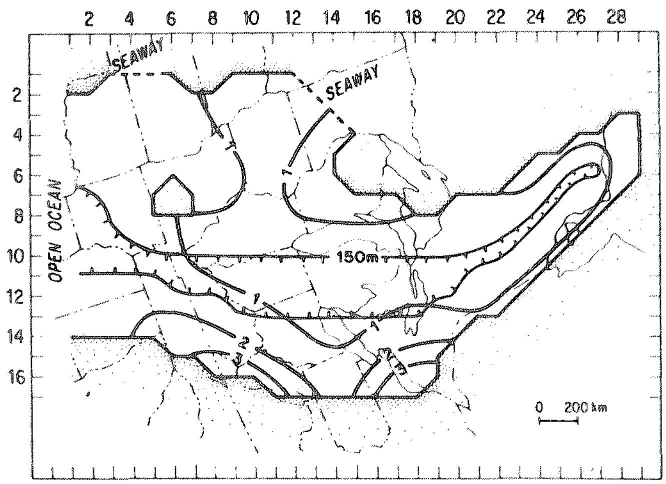


Figure 1. Paleotidal ranges (dark line in meters) in the Late Devonian Catskill Sea of eastern North America as predicted by the hydrodynamic model. Basin geometry is modified from Heckel and Witzke (1979) with paleonorth to the top and the center of the sea at approximately 10°S latitude. The 150 m bathymetric contour encompasses the Upper Devonian black shale facies; water depth decreases linearly landward from that contour. Open ocean tidal range at column 1 is 1 m, today's average ocean tidal range near the edge of the world's continental shelves. Tidal augmentation occurs along the Middle Atlantic States.

To address the second question, the same hydrodynamic model was modified to calculate resonance effects on generic margins of rectangular cross section and unit length alongshore (Slingerland, 1985). The forcing tide at the margin edge was a simple sinusoidally varying astronomical tide of constant amplitude equal to 0.45 m; the shore was a perfect reflector of the tidal wave. Thirty-eight numerical experiments were performed using various margin widths and friction factors. The results pertinent to this discussion are presented in Figure 2.

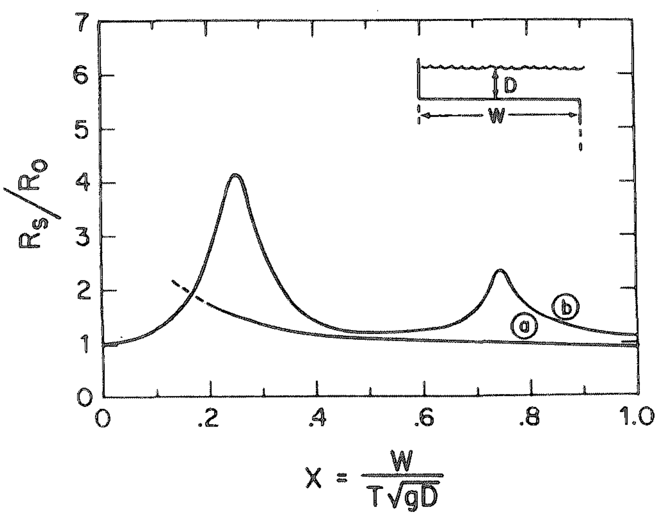


Figure 2. The relationship between dimensionless tidal range at the shore and dimensionless shelf width as predicted by the hydrodynamic model. The shelf is of unit alongshore length and uniform water depth D, equal to 20 m for curve (a) and 60 m for curve (b). Manning's n for both curves is 0.04. R_s (m) = computed tidal range at the shore, R_o (m) = given ocean tidal range at the margin edge (0.45 m), W = shelf width (m), T = M_2 tidal period = 44700 secs, and g = gravitational acceleration. Tidal augmentation due to resonance becomes important at some water depth between 20 and 60 m. Maximum augmentation occurs at quarter multiples of the tidal wavelength.

Discussion and Conclusions

Figure 1 depicts the modelled basin geometry and resulting co-range lines for the most probable boundary and initial conditions. Remarkably, the tides are augmented in this epicontinental sea by up to three times the ocean range with the highest values occurring along the Middle Atlantic States. Other experiments (Slingerland, 1986) have shown that the results are relatively insensitive to the bathymetry of the basin floor and to the open ocean tidal range, but are reduced as the entrance to the seaway is narrowed or the concavity of the southern (Devonian coordinates) shoreline is decreased. Thus, the answer to the first question is yes, mesotides (1-3.5 m range) are theoretically possible on the Catskill basin margin given our present best estimates of paleogeography.

The answer to the second question depends upon the causes of the augmentation. In general, augmentation of long waves is due to convergence and resonance. By convergence is meant a shoreward decrease in margin water depth or decrease in alongshore crest length of the tidal wave. For a frictionless, nonreflecting wave it is easy to show that if the wave energy is transmitted undiminished, then the amount of augmentation due to convergence is proportional to the quarter root of the convergence due to depth changes and the square root of the convergence due to crest length changes (Ippen, 1966, p. 502). Thus, a tidal wave traveling towards the Acadian Highlands across what is Pennsylvania today (Fig. 1) would have experienced approximately a threefold decrease in water depth and a one and one half-fold decrease in crest length, and therefore would have undergone a little greater than three-halves augmentation in height. That is approximately half the augumentation calculated in the model. This is only a first-order estimate because the actual tidal wave in the Catskill Sea propagated obliquely to the Acadian margin.

The other cause of augmentation is resonance, an oscillation of waters on the basin margin in phase with the forcing tides at the margin's edge. Figure 2 shows the amount of resonance augmentation predicted by the hydrodynamic model as a function of dimensionless margin width. Even at the high rates of frictional energy loss used in the model and a shallow depth (20 m), augmentation due to resonance occurs for dimensionless widths up to 0.9 (about 563 km for an M_2 tidal wave). For deeper waters (60 m), resonance augmentation occurs for all dimensionless widths up to 1 (about 1084 km) with two maxima at widths that are 1/4 and 3/4 multiples of the tidal wavelength.

The resonance portion of the augmentation seen in Figure 1 may be estimated from Figure 2. If the margin width through Pennsylvania is approximately 400 km (Fig. 1) and the average depth is 75 m, then the dimensionless width is 0.33 and R_S is greater than 2 times R_0. The two analyses taken together suggest that the predicted augmentation on the Catskill margin in Figure 1 is due about equally to convergence and resonance. Thus, decreasing the concavity of the shoreline in Figure 1 should reduce the augmentation by up to one-half. Increasing the dimensionless width of the basin margin (as in a transgression) from 0.33 while keeping the depth equal to 60 m would drastically decrease the resonance augmentation at first (Fig. 2), whereas if the depth were 20 m, increasing the dimensionless width would only slightly decrease the augmentation.

In conclusion, tidal flows could have played a more significant role in distributing sediment on the Catskill margin than previously recognized. Their contribution relative to storm-driven flows would have varied as a function of the local width, depth, and concavity of the shoreline, and therefore possibly with transgressions and regressions, an idea presently being considered by John S. Bridge (personal communication). A logical next step is to combine facies data like those in Table 1 with model results to determine the paleogeographic locations and ages in which each process was dominant.

REGIONAL DISTRIBUTION OF FACIES IN THE CATSKILL FORMATION AND THE CONTROLS ON RED-BED COPPER-URANIUM OCCURRENCES

by
Arthur W. Rose, Arthur T. Smith, and Christopher H. Gammons

Introduction

The studies of Rahmanian (1979) in central Pennsylvania as summarized by Williams in this volume clearly show the existence of several facies of Catskill sedimentation, and the existence of a large input center that remained in the vicinity of Port Matilda through Catskill time. Various other workers have identified facies, members, and other units in the Catskill over limited areas (Willard, 1939; Allen and Friend, 1968; Walker and Harms, 1971; Glaeser, 1974; Humphreys and Friedman, 1975; Epstein and others, 1974; Dyson, 1963, 1967), but a regional evaluation of facies has been lacking. As part of a regional study of red-bed copper-uranium occurrences, Smith (1983) and Smith and Rose (1985) have extended Rahmanian's work to furnish a regional correlation scheme and attempted to show how the groups of facies are interrelated within Pennsylvania. This paper summarizes this regional pattern, and also describes the relation of the copper-uranium occurrences to regional and local sedimentary characteristics.

Regional Distribution of Facies

Most previous work on the Catskill has concentrated on sedimentary facies within relatively limited areas of a few quadrangles. Stratigraphic units within these areas have been defined as members (Irish Valley, Sherman Creek, Duncannon, Clarks Ferry, Walcksville, Long Run, etc.), and correlated over distances of a few tens of kilometers. Over larger distances, especially east-west, these units seem to grade into each other, and correlation has not been attempted.

The approach of Smith and Rose (1985) was to define groups of facies, termed magnafacies, based on certain sequences of facies, and to correlate these magnafacies in widely spaced stratigraphic sections. Table 1 summarizes the characteristics used to define the facies. Associations of these 10 facies were then used to define four magnafacies, as illustrated on Figure 1. These magnafacies show consistent regional relations.

Magnafacies A generally forms a basal unit of the Catskill. It is composed dominantly of interbedded shale and fine sandstone, usually including both gray and red units, and is interpreted to have been deposited in an alternating marine to nonmarine tidal-flat environment. This magnafacies includes the Irish Valley Member in the field trip area, as well as the Towamensing and Beaverdam Run Members along the Lehigh River.

Magnafacies B is dominated by thick red shales accompanied by thin fine-grained sandstones, commonly in fining-upward cycles, and is interpreted as a low-energy fluvial deposit predominantly formed on relatively inactive parts of the coastal plain. Occasional thin sandstones interpreted as being of transgressive, tidal origin also occur in Magnafacies B. Magnafacies B is represented by the Sherman Creek Member in the field trip area, and by the Walcksville and Long Run Members in the Lehigh River area.

Table 1. Generalized criteria which define 10 facies associations in the Catskill Formation of Pennsylvania (after Smith and Rose, 1985).

Facies	Color[1]	Fossils[2]	Geometry	Grain size and lithology[3]	Sedimentary[4] structures	Commonly associated features
1	g	M and/or P	tabular	f-vf ss	massive beds, flasers, ripples, trough x-beds	--
2	g	M and/or P	lenticular	f-c ss ± quartz pebbles	do.	--
3	g	M and/or P	tabular	do.	do.	--
4	r or g	P	lenticular	f-vf ss	trough and planar x-beds, erosive basal contacts	fines up, basal cgl lag
5	g or r	P	do.	m-c ss	do.	cgl lag
6	g	P	do.	cgl	massive	facies 4, 5
7	g	M and/or P	lent./tabular	stst, mdst	bioturbated	thin, fss and facies 1, 2, 3
8	g	P	do.	f-c stst	bioturbated, root casts	do.
9	r	-	do.	stst, mdst	root casts	fines up and underlain by facies 4
10	g	P	do.	stst, mdst	--	facies 5

[1]g = gray; r = red (first letter is most common).
[2]M = marine; P = fossil plant material.
[3]f = fine; vf = very fine; c = coarse; m = medium; cgl = conglomerate; ss = sandstone; stst = siltstone; mdst = mudstone.
[4]x-beds = crossbeds.

Magnafacies C is composed dominantly of fine- to coarse-grained sandstone and conglomerate, usually gray, with minor shale. These thick sandstone units are interpreted as the product of large braided rivers carrying abundant coarse detritus from the uplifted source area to the east. Magnafacies C forms three northwest-trending lobes in Pennsylvania and southern New York (Figure 2A). These lobes are in the Harrisburg-Port Matilda zone, the Lehighton-Scranton zone, and southern New York. Part of the "Duncannon" Member visited in the Port Matilda section of the field trip belongs to Magnafacies C, as does the Clarks Ferry Member along the Lehigh River.

Magnafacies D is composed of thick fining-upward cycles with subequal amounts of gray or red sandstone and red shale. These sediments are inferred to have been deposited by meandering rivers and are thickest in the same areas that Magnafacies C is thick (Fig. 2B). Magnafacies D corresponds to the Duncannon Member in the sections along the Lehigh River, the lower Juniata Valley (Dyson, 1963, 1967), and the I-80 section of the field trip, and to part of the Duncannon section at Port Matilda.

The isopachs for Magnafacies C and D show a clear lobate zone extending northwest from the Lehigh River area, and a second thick lobe extending from Peters Mountain along the Susquehanna River northeast to the Port Matilda section visited on the field trip. A third area of thick sediment is suggested in southern New York state. Between these lobes, Magnafacies C disappears and Magnafacies D thins or disappears. This pattern is indicated on section A-A' on Figure 3. In general, the magnafacies tend to thin to the northwest, B and C most markedly (Fig. 3).

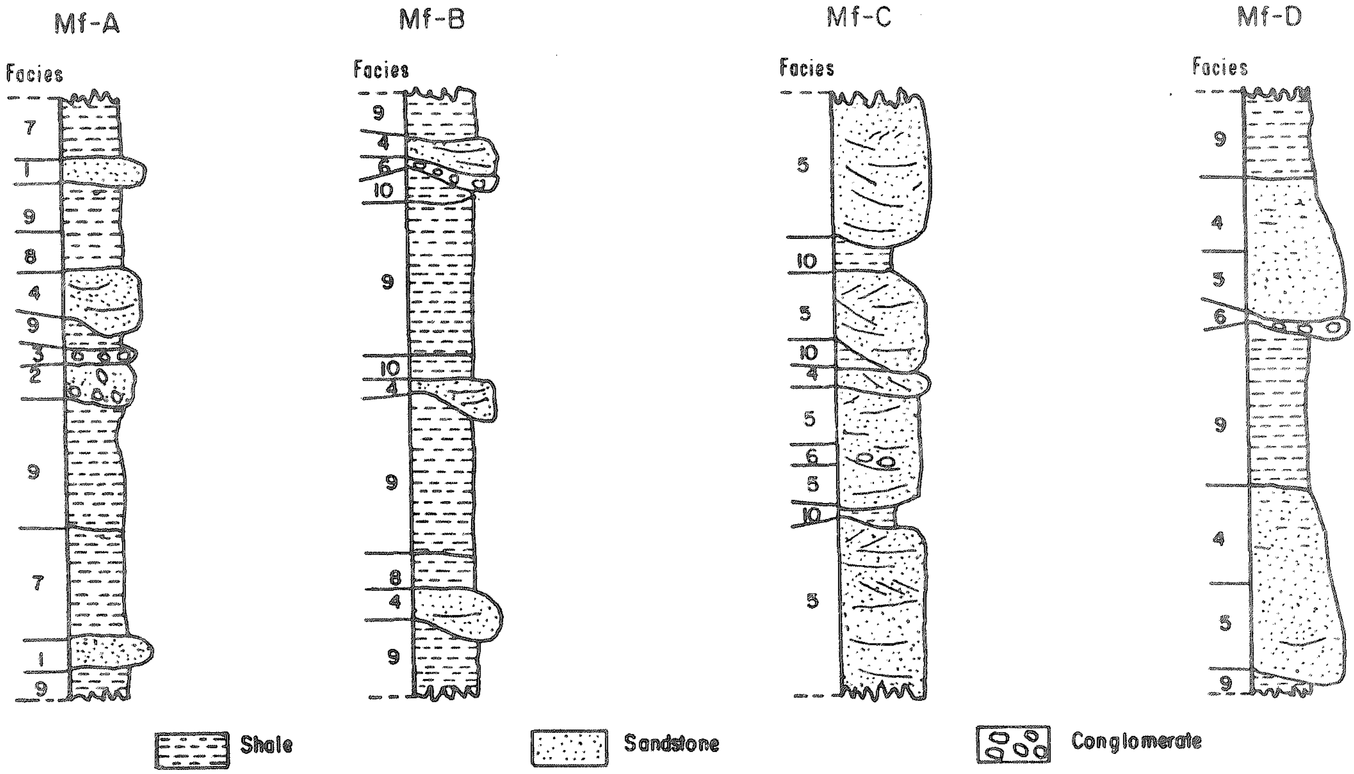


Figure 1. Idealized stratigraphic sections for four magnafacies (Mf) used in regional correlations of the Catskill Formation (Smith and Rose, 1985). See Table 1 for descriptions of component facies. Not to scale.

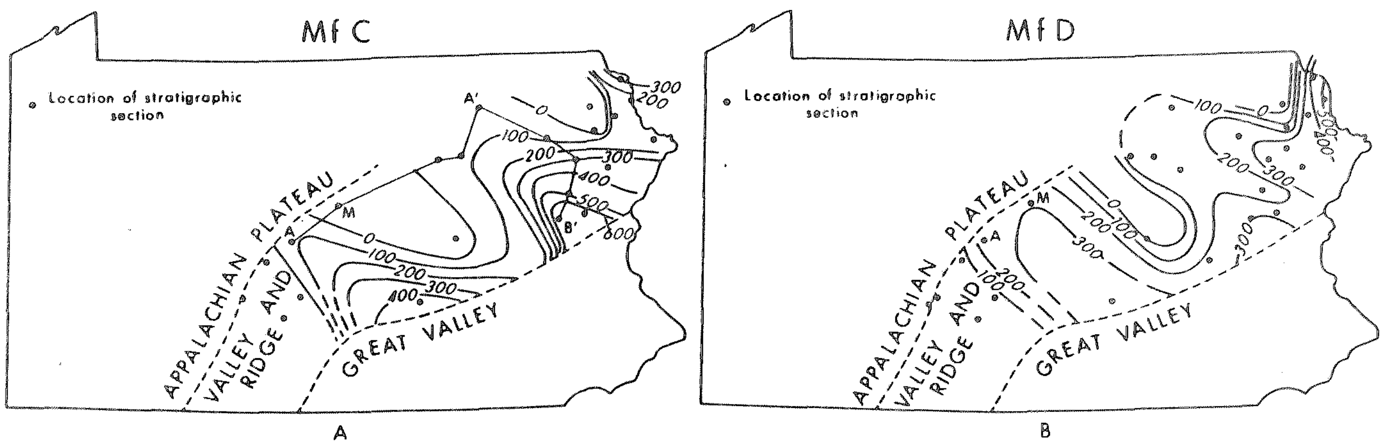


Figure 2. Isopach maps of Magnafacies C and D, in meters, showing lobes of thick sediment in the Port Matilda, Lehighton, and Southern New York areas. Sections A-A' and A'-B' are on Figure 3. Point A is the Port Matilda section and point M is the Milesburg (I-80) section. Modified from Smith and Rose (1985) as described in Appendix.

Given the coarse fluvial nature of most sediment in Magnafacies C and D, these lobes appear to be the loci of major river systems that crossed the alluvial plain from the source area to the east.

Red-Bed Copper-Uranium Occurrences

At least 80 small occurrences of copper/uranium are known in the Catskill Formation in Pennsylvania (Smith and Hoff, 1984; McCauley, 1961; Smith, 1980, 1983;

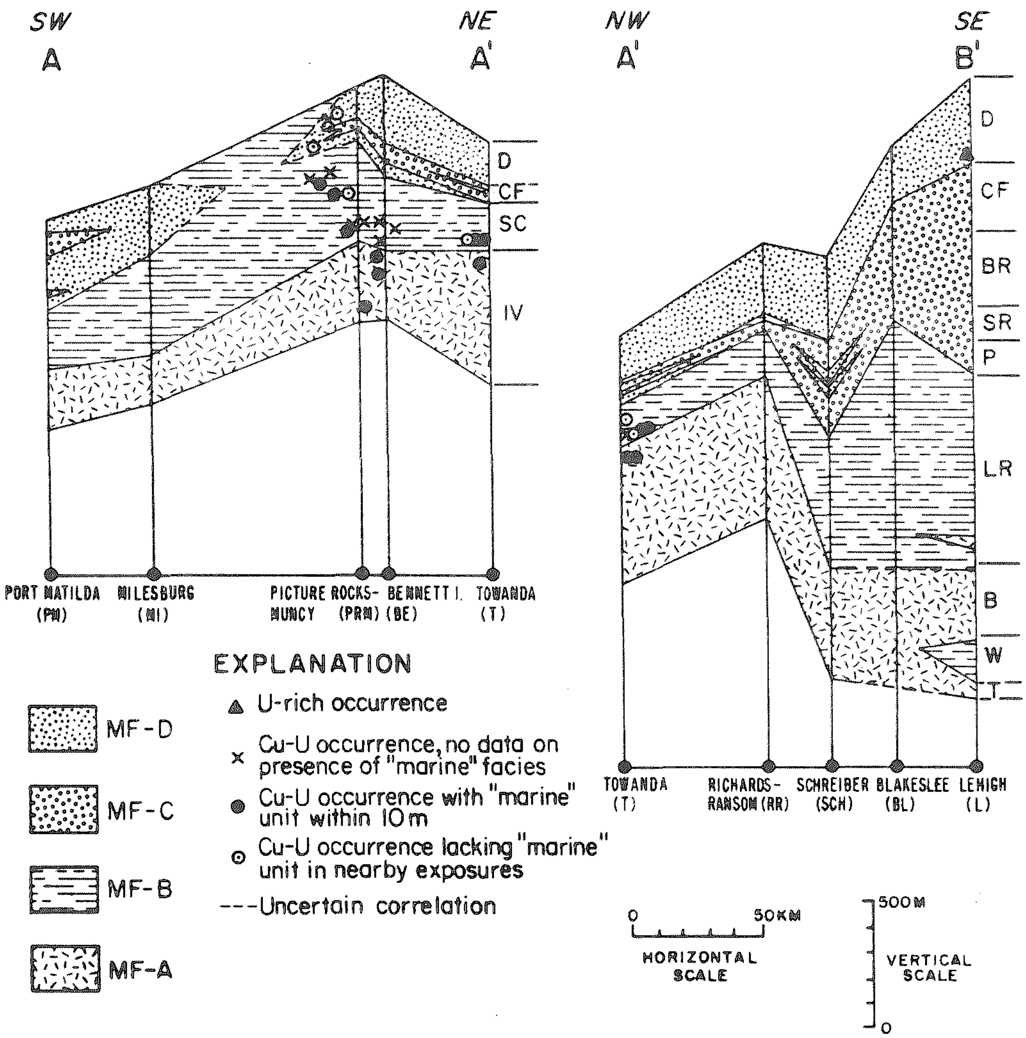


Figure 3. Regional cross sections showing distribution of magnafacies and Cu-U occurrences (modified from Smith and Rose, 1985, as described in the Appendix). Datum is Tioga bentonite +1500 m; nonpatterned area is marine Devonian. Abbreviations: IV, Irish Valley; SC, Sherman Creek; CF, Clarks Ferry; D, Duncannon; T, Towamensing Member; W, Walcksville Member; B, Beaverdam Run Member; LR, Long Run Member; P, Packerton Member; SR, Sawmill Run Member; BR, Berry Run Member. Slightly modified from Smith and Rose (1985).

Klemic, 1962; Sevon and others, 1978). These occurrences have the following general characteristics:

1. Mineralization is associated with and commonly replaces fossil plant material within gray or green facies of the Catskill Formation.

2. Mineralization is discontinuous but tends to be lensoid and stratiform. Typical mineralized zones are 1-50 cm in thickness and 1-10 m in length along bedding. Copper concentration is in the range of 0.05 to 0.3% Cu, and uranium contents are usually a few tens of parts per million. None of the presently known occurrences appear promising for economic development, though two produced small amounts of ore in the past.

3. The primary copper minerals appear to have been chalcocite and similar minerals, but partial to complete oxidation to bright-green malachite is common. Uranium occurs as uraninite at a few localities, but may be present in the organic matter at others. Many other minerals have been identified locally (Smith and Hoff, 1984).

4. Host rocks for mineralization are most commonly shales, but some are sandstones or conglomerates containing caliche pebbles and plant trash at the base of fining-upward cycles.

5. Features which suggest that the occurrences formed at low temperatures (<100°C) include orthorhombic chalcocite, exsolution of bornite from chalcopyrite on heating above 75°C, a wide range in $\delta^{34}S$ of sulfides, an apparent precompaction age relative to diagenetic minerals, fine grain sizes, and a general lack of evidence for elevated temperatures in the vicinity (McCauley, 1961; Smith, 1983; Rose and others, in press).

6. Sandstone beds with inferred original permeability higher than average for the Catskill are usually present within a few meters stratigraphically (Smith, 1983), and many occurrences are in zones of Magnafacies B containing nearby tongues of tidal sandstone.

7. Occurrences are most abundant within Magnafacies B, near the margins of large lobes of Magnafacies C and D (Fig. 4). A few occurrences are in the upper part of Magnafacies A and in Magnafacies D. None are in Magnafacies C, though this is the site of the uranium occurrences (no copper) near Jim Thorpe in eastern Pennsylvania.

Based on these characteristics, we infer that Cu and U were emplaced during diagenesis by pore fluids moving preferentially along the more permeable sand bodies in the Catskill. The copper and uranium were precipitated by the reducing effect of plant fragments and the accompanying bacterial sulfate reduction. Copper is not easily mobilized under oxidizing conditions except at acid pH, but in chloride-bearing waters at intermediate oxidation states it is very soluble as cuprous chloride complexes (Rose, 1976). The chloride may have been supplied to the pore waters by the occasional marine transgressions, or by evaporation of fresh waters in the relatively arid red-bed environment. Uranium is soluble under oxidizing conditions, including those under which cuprous chloride complexes are stable. Uranium mobility is accentuated at higher pH values, especially if dissolved carbonate is high, as is common in arid environments.

The source of the Cu appears to be the red beds themselves. Figure 5 shows that Cu content of pre-Catskill detritus, as indicated by the Cu content of marine rocks beneath the Catskill, was 20-25 ppm. No major change in composition of detrital material is expected for the subsequent nonmarine Catskill sediments, yet typical Cu contents of red beds are less than 5 ppm (Fig. 5). In contrast, green and gray Catskill sediments contain a wide range of Cu contents, up to hundreds of parts per million. A major redistribution of Cu has evidently occurred in the Catskill, leading to mobilization of copper out of the red units into some of the reduced (gray-green) units.

Data for Zn, Ni, and Co in the same samples show no differences between pre-Catskill and Catskill, and generally only small changes for Pb. These elements, unlike Cu and U, are not concentrated appreciably in the red-bed mineralization, and do not show depletion from the red beds.

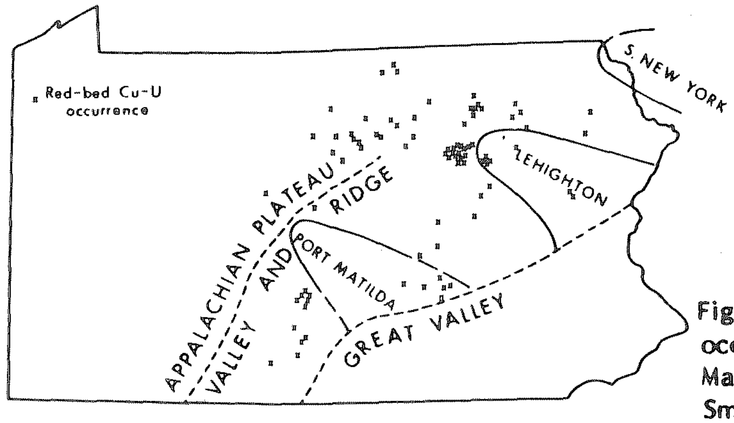


Figure 4. Location of red-bed copper-uranium occurrences relative to lobes of thick Magnafacies C. Modified from Smith and Rose (1985).

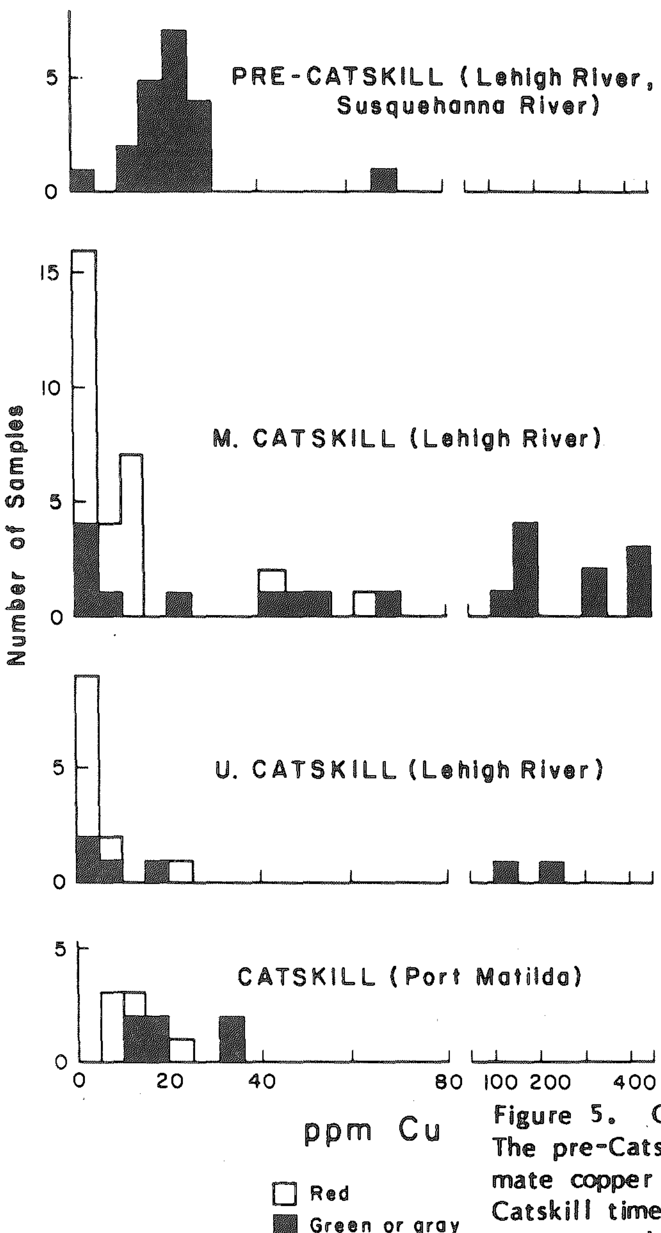


Figure 5. Copper content of Catskill and pre-Catskill rocks. The pre-Catskill rocks are inferred to indicate the approximate copper content of detritus entering the basin during Catskill time. The red beds have generally been depleted in copper, and green-gray beds have commonly been enriched.

Data for Cu content of samples from the Port Matilda section are also shown on Figure 5. Though the effects are not as extreme, the same tendency for depletion of Cu from red beds and enrichment in gray beds is noted.

One small copper occurrence has been found in the Port Matilda section, near the base of the exposures examined in the Irish Valley Member. It appears that Catskill sediments of Magnafacies B in the Port Matilda section may have lacked reduced zones or that chloride content remained low because of underflow of fresh groundwater down the alluvial plain beneath the river system at this point, thus leading to only rare accumulations of Cu.

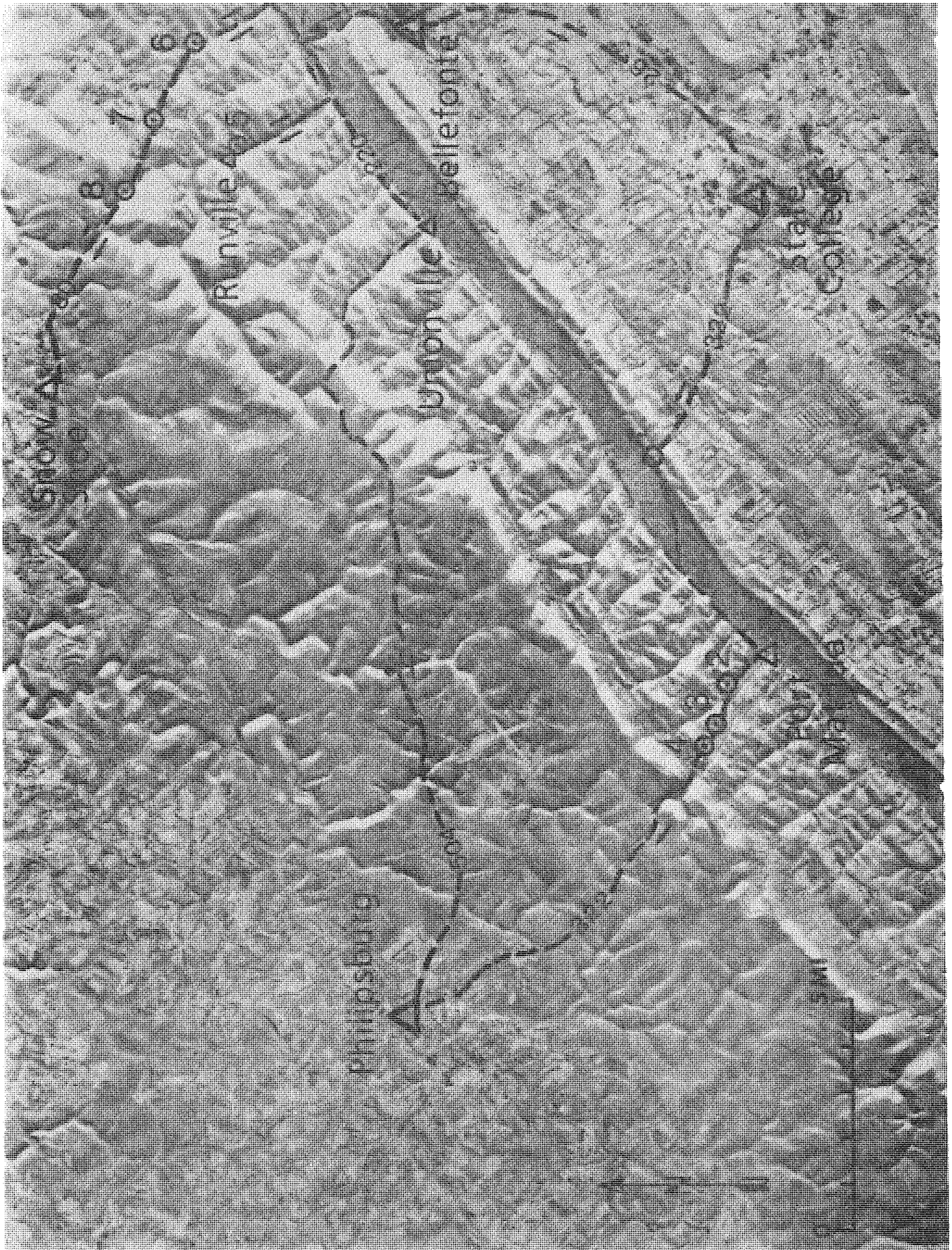
Appendix

Data on thickness of Magnafacies C and D in addition to that reported by Smith and Rose (1985) and Smith (1983), and shown on Figure 2, are as follows:

Location of Section	Mf C	Mf D
Raystown Dam	0 m	< 50 m
Entriken	0	< 100
Horseshoe Curve/ New U.S. 22	0	< 50
Tyrone (Pa. 453)	0	< 100

These data demonstrate a rapid southwestward thinning of the Mf C and Mf D lobe at Port Matilda. Although exposures are poor, no more than one cycle of Mf D appears to exist at these localities.

In addition, a reexamination of the Port Matilda section indicates that an additional 56 m of the section fits the criteria for Mf C, with correspondingly less Mf D.



Stop localities for Field Trip #2, plotted on Radar Map of part of Central Pennsylvania (STAR-1 Imagery courtesy of INTERA Technologies Inc., September, 1984).

FIELD TRIP #2

FIELD GUIDE - CATSKILL SEDIMENTATION IN CENTRAL PENNSYLVANIA

E. G. Williams and Rudy Slingerland,
Field Leaders

Introduction

This field trip will attempt to illustrate the conclusion, presented by Rahmanian (1979) and Williams (this volume) that Catskill sedimentation in central Pennsylvania occurred in tide-dominated deltaic and interdeltaic environments, the latter exhibiting features of an open as well as a barred coastline. Specifically, the tide-dominated delta occupied an area extending approximately from Tyrone on the south to midway between Port Matilda and Milesburg on the north, a distance of about 30 miles. The rocks recording its existence are the sandstone-shale facies of the upper Trimmers Rock Formation, a noncyclic facies of the Irish Valley Member of the Catskill Formation, and a silty facies of the Sherman Creek Member of the Catskill Formation. Reference to Figure 1 shows that to the north, at Milesburg, the sandstone-shale facies of the Trimmers Rock Formation is absent and the motif or cyclic facies of the Irish Valley Member replaces the nonmotif facies of the Port Matilda section. Well-developed fining-upward cycles of red sandstone and mudstone of the Sherman Creek Member at Milesburg replace the finer grained noncyclic facies present at Port Matilda. In addition, but not shown on the diagram, important facies changes occur in both the Duncannon Member of the Catskill Formation and the Pocono Formation between Milesburg and Port Matilda, namely the fining-upward alluvial cycles become much thicker, fewer in number, and coarser grained at Port Matilda. Rahmanian and Williams interpret all the above facies changes, which also occur to the south of Tyrone, to be the result of the presence of a major depocenter whose axis was located at Port Matilda. There, a large braided river (represented by the Pocono Formation) supplied sediments to a delta consisting of anastomosing distributaries on the upper part (Duncannon Member) and tidally influenced channels, mud flats, shallow bays, and bars on the lower parts of the delta (Sherman Creek and Irish Valley Members). These graded seaward into storm-deposited shelf deposits (upper Trimmers Rock Formation). To the north and south of the delta existed broad muddy tidal flats, developed along both barred and open coastlines (motif facies of the Irish Valley Member), landward of which was a wide coastal plain crossed by low-gradient, meandering rivers (cyclic facies of the Sherman Creek and Duncannon Members).

The strategy of this trip will be to compare facies within and between the various members of the Trimmers Rock and Catskill Formations at the Milesburg and Port Matilda sections. The stops and the general facies and members seen at each are summarized in the following table:

Port Matilda transect (Route 322)

Stop	Description
1	Overview at Skytop.
2	Sandstone-shale facies and silty shale facies of upper Trimmers Rock Fm.
3	Nonmotif facies of Irish Valley Mbr., Catskill Fm.
4	Complex cycles of Duncannon Mbr., Catskill Fm.

Milesburg transect (Route I-80)

Stop	Description
5	Motif facies of Irish Valley Mbr., Catskill Fm. (Runville, Pa.)
6	Silty shale facies of upper Trimmers Rock Fm.
7	Cyclic mudstone-sandstone facies of Sherman Creek Fm.
8	Simple cycles of Duncannon Mbr., Catskill Fm.

The interval to be examined at each locality is noted on the appropriate stratigraphic section on Plate 1.

INC. MIL.	CUM. MIL.	DESCRIPTION
0.0	0.0	LEAVE parking lot of Holiday Inn, State College, Pa. Turn left (west) on Route 322 towards State College.
3.5	3.5	The red soils and scrub oak forests seen on both sides of the road are characteristic of the Upper Cambrian Gatesburg Formation, a cyclic orthoquartzite-dolomite on which very thick, kaolinitic, hematitic soils have developed during the late Tertiary. The soils are sandy and consequently the region underlain by the Gatesburg Formation is underdrained, producing a terrain called "The Barrens." The soils have been mined in the past as a source of iron ore and refractory clays; the iron mining town of Scotia, southwest of here, was founded at the turn of this century by Andrew Carnegie and named after his native Scotland.
2.85	6.35	Outcrop on right of Upper Cambrian Warrior Formation, the oldest exposed formation in the Nittany Valley. The small valley 100 m ahead (to the west) is a manifestation of the Birmingham thrust fault which brings the Warrior Formation in contact with the Bellefonte Dolomite, outcrops of which form the western valley wall.
1.25	7.60	Ahead, the road ascends Bald Eagle ridge, the westernmost ridge of the Appalachian Mountains in this area. Rocks exposed on the ascent are black silty shales and sandstones of the Upper Ordovician Reedsville Formation, a structurally thinned section of sandstones of the Bald Eagle Formation forming a secondary ridge, red sandstones, siltstones, and shale of the Juniata Formation of lesser resistance, and orthoquartzites of the Lower Silurian Tuscarora Formation forming the main ridge. The beds are vertical to slightly overturned.
0.85	8.45	**Stop 1: Skytop Overlook.** Discussant: E. G. Williams.

This locality sits on the crest of Bald Eagle Mountain overlooking Bald Eagle Valley and the Allegheny escarpment to the northwest. The valley separates the gently dipping upper Paleozoic rocks of the Allegheny Plateau to the west from the folded rocks of the Valley and Ridge to the east (see Gold, this volume, for details). Bald Eagle Mountain, which has an elevation of 2300 feet, is underlain by the Tuscarora Formation, a resistant orthoquartzite of Early Silurian age. The topmost and steepest slope of the escarpment is formed by the Pocono Formation, a low-rank graywacke, which has a quartz content of about 80 percent and is believed

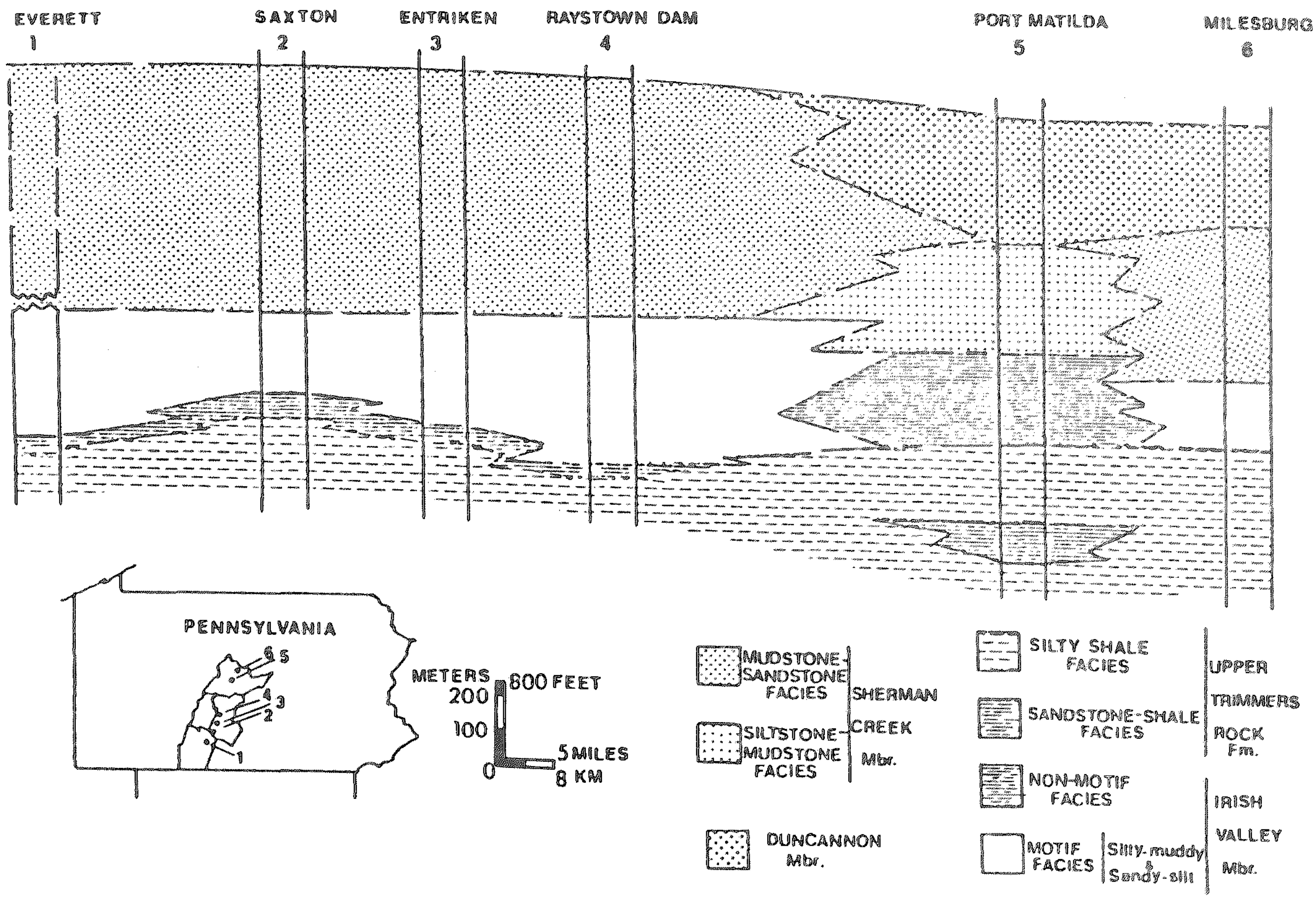


Figure 1. Generalized lithofacies cross section of the Upper Devonian rocks in central and south-central Pennsylvania (from Rahmanian, 1979).

to represent deposits of a braided river. The strike valley below the Pocono marks the top of the Catskill Formation, which is divided into three members, from bottom to top, the Irish Valley, Sherman Creek, and Duncannon. The several ridges below the steep slope of the escarpment are sandstones of the Duncannon Member, each 50 to 100 feet thick, inferred to be migrating alluvial channels which supplied sediment to the Catskill delta manifested in the ridges and strike valleys in the middle part of the escarpment. These constitute the Irish Valley Member. The ridges are distributary-mouth bars or tidal-sand ridges, and the strike valleys represent bay and tidal-flat muds. The several ridges in the lower part of the escarpment are underlain by shelf sandstones of the upper third of the Trimmers Rock Formation. All of the formations visible in the Allegheny escarpment intertongue and so, on a regional basis, are equivalent in age.

The gently undulating surface on the skyline is the Schooley peneplain of Johnson, an early Tertiary erosion surface. Based on measurements of porosity and bulk density of Pennsylvanian sandstones and reflectance of coals, an estimated 15,000 feet of denudation has occurred at the Allegheny escarpment since the Permian (Paxton, 1983).

LEAVE STOP 1 and proceed west on Route 322.

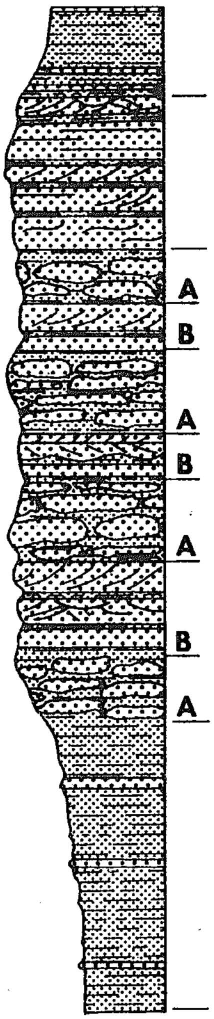
2.45	10.90	STOP SIGN; TURN LEFT on Route 220 South/322 West.

1.50 12.40 Outcrop on right of Upper Devonian Harrell Shale.

1.40 13.80 STOP LIGHT in Port Matilda; TURN RIGHT, continuing west on Route 322.

1.80 15.60 **Stop 2: Upper part of the Trimmers Rock Formation.** Outcrops are along a small gravel side-road. Discussants: E. G. Williams and R. Slingerland.

The rocks exposed at this locality comprise the silty shale and sandstone-shale facies of the upper part of the Trimmers Rock Formation. The sequence correlates with the exposure of the silty shale facies at Stop 6 (refer to Plate 1, section 7, 0-400 ft interval). A summary of the major lithologies and their environmental interpretations is presented in Figure 2. The lower 75 feet of the outcrop resembles the silty shale facies described at the Milesburg section. The sandstone-shale facies here displays entirely different sedimentary characteristics than have been observed at any of the other sections. It contains two different sandstone



Facies	*Sedimentary structures*	*Interpretation*
Brown-gray fossiliferous fine-grained sandstone	Horizontal, flaser, lenticular, and large- and small-scale, planar and trough crossbedding	Tidal sand bar (low-tide terrace?)
Sandstone-shale facies A: Brown, massive to thickly bedded, fine- to very fine grained sandstone and siltstone B: Olive-green, fine-grained, fossiliferous, regularly bedded sandstone; silty shale and shale	– A: Abundant soft-sediment deformational structures; well-preserved horizontal lamination – B: Hummocky cross-strata; wavy, flaser, and rhythmic bedding; clay drapes on foresets; ripple lamination; bioturbated horizons common	Alternation of destructive (marine-influenced) and constructive (fluvial-influenced) phases of sedimentation in a storm-influenced prodelta and delta-front environment
Silty shale facies Gray olive-gray, fossiliferous silty shale to siltstone with thin sandstone layers interspersed	– Siltstones: Dominantly horizontally laminated – Sandstone: Small- to micro-scale cross-bedding, linear and interference ripples; hummocky cross-stratification	Shallow marine shelf

Figure 2. Idealized vertical facies sequence of the uppermost Trimmers Rock Formation from pooled data at the Port Matilda section (modified from Rahmanian, 1979).

subfacies that occur together with interlayered siltstone and silty shale units in a semicyclic arrangement. These are called subfacies A and B (Fig. 3).

Subfacies A consists of chocolate-brown, thickly bedded and sometimes massive, very fine grained, micaceous sandstone and chocolate-brown and thinly bedded sandy siltstone, occurring in units 8-18 feet thick. A characteristic aspect of this facies is the large number of soft-sediment deformation features such as ball-and-pillow structures and convoluted laminations. Marine fossils are very rare and burrows and bioturbated sediments are absent.

Sandstone subfacies B, which alternates with subfacies A, consists of an alternation of light-gray sandstone and silty shale units which are devoid of any soft-sediment deformation features. The sandstone units have a higher quartz content and are coarser grained than those in subfacies A. Thicker beds are

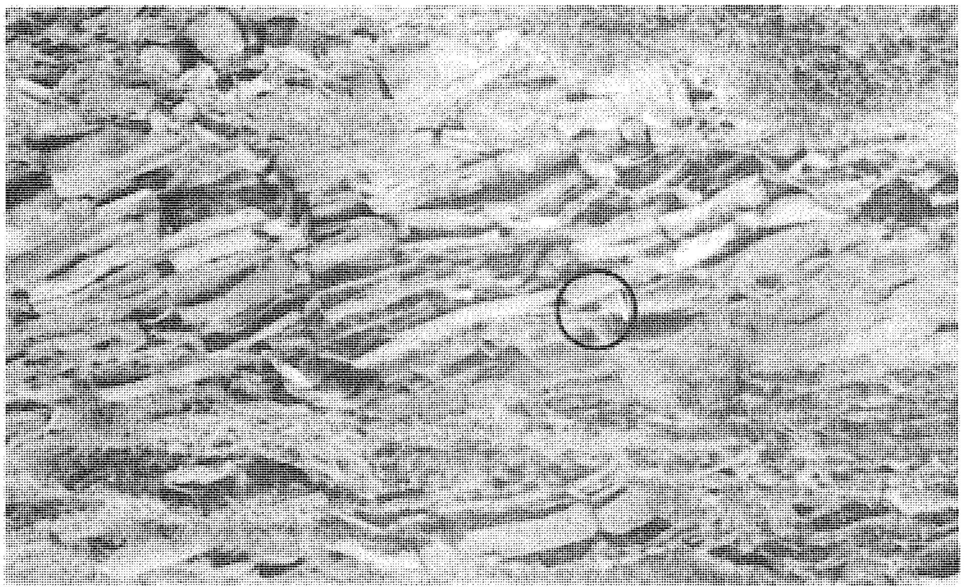


Figure 3. Alternation of sandstone subfacies A (above the hammer) and subfacies B (below) in the uppermost Trimmers Rock Formation at Stop 1.

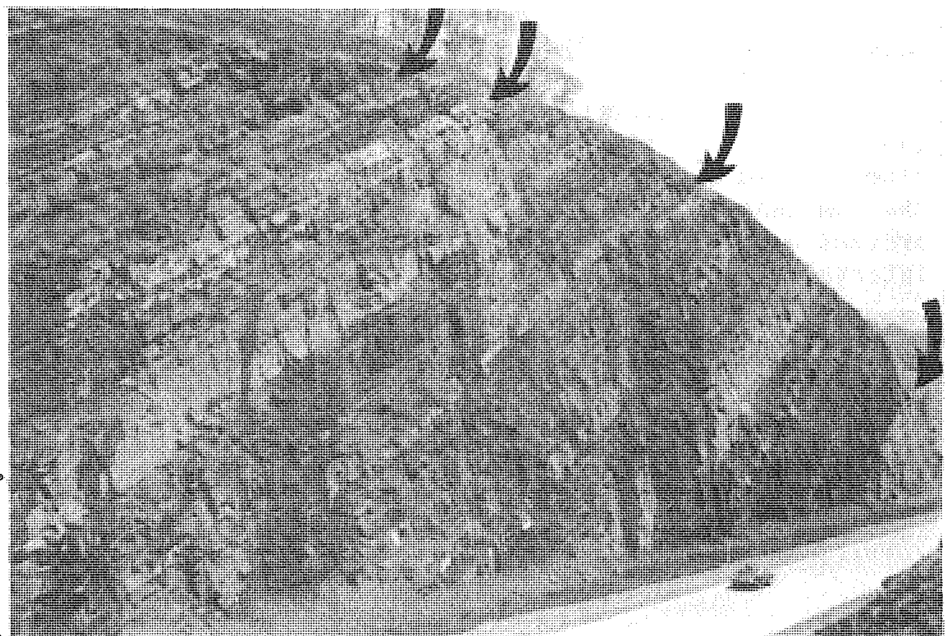


Figure 4. Exposures of the Irish Valley Member of the Catskill Formation at Stop 3. Arrows point to marine horizons. Note the anomalously high sand/shale ratio compared to Irish Valley sections to the north and south.

commonly both horizontally and cross-laminated; thinner beds exhibit rippled upper surfaces with clay drapes. Very thin beds exhibit flaser and lenticular bedding. Interbedded siltstones and shales are extensively burrowed. Marine fossils are abundant and scattered throughout this facies, but are most commonly found as brachiopod and coquina layers at the tops and bottoms of sandstone beds.

The overall characteristics of this facies are suggestive of a shallow marine prodelta-subtidal environment. The presence in subfacies B of flaser and lenticular bedding, hummocky cross-lamination, clean, well-sorted sandstones, clay drapes over rippled surfaces, extensive bioturbation, and abundant marine fossils points to deposition in a subtidal environment, characterized by alternation of storm-current bed-load sediment transport with deposition of the suspended load during slack water periods. Subfacies A displays features suggesting rapid sedimentation in a low-energy environment characterized by lower flow energies. The presence of ball-and-pillow structures, the absence of bioturbation structures and burrows, and the lack of marine fossils are taken together to mean sedimentation was rapid. Immature texture and composition of sandstones of this facies suggests little, if any, reworking by tides and waves. These features are suggestive of fluvially induced sedimentary processes in a prodelta, delta-front environment. It is suggested that these sediments were deposited basinward from the delta mouth in the absence of significant marine reworking. As soon as the rate of sediment supply was reduced, either as a result of lateral shift of the mouth or seasonal variations in river discharge, marine processes took over. During this phase of sedimentation, sediments were reworked and redistributed by waves and tidal currents into various types of shallow marine subtidal sandstone bodies, eventually producing subfacies B. Repeated alternation of the two subfacies might result from the process of delta switching.

LEAVE STOP 2 and proceed west on Route 322.

0.80 16.40 Stop 3: Irish Valley Member of the Catskill Formation. Discussants: E. G. Williams and R. Slingerland.

The rocks exposed at this locality (Fig. 4) are part of the nonmotif facies of the Irish Valley Member of the Catskill Formation and are correlated with those seen at Stop 5 (Plate 1, compare section 7, 1450-1700 ft, to section 6, 0-200 ft). There are two facies exposed; these are, from base to top: 1) green quartzitic sandstone-mudstone facies; and 2) interbedded red sandstone, siltstone, and mudstone facies.

The green quartzitic sandstone-mudstone facies in the lower part of the outcrop consists of a vertical alternation of thinly bedded olive-green mudstone and very fine grained quartzitic sandstone and siltstone. Wavy and lenticular bedding are the dominant sedimentary structures. Small brachiopods and bone fragments are present but uncommon. In the lowermost parts of the exposure occur two thin intervals of red, fissile, Lingula-bearing shale and siltstone.

The red facies consists of fine- to medium-grained, well-sorted sandstone beds interbedded with red, rooted and burrowed, laminated to massive mudstones and siltstones. Sandstones occur in two distinct groups of thickness--large (14-19 feet) and small (up to 4 feet). The larger units have relatively sharp and flat but generally interfingering basal contacts, with large-scale trough and planar crossbedding in the lower parts which grade upward to ripple and micro cross-lamination at the top. All measured crossbed dip directions are to the northwest (Williams, this volume, Fig. 8), but some inaccessible beds possess apparent dip directions to the southwest. Two of the large sand bodies are completely cut by large channels trending approximately southwest-northeast, and filled with siltstone. At the tops of the sandstones and at other positions within the deposit occur thin, green, rippled sandstone and siltstone containing marine fossils.

The red mudstone and siltstone are arranged into coarsening-upward cycles, beginning with massive, burrowed mudstone which grades upward to interbedded red siltstone and sandstone beds. The top of the cycle may consist of very fine, rippled, green quartzitic sandstone, often with marine fossils.

The assemblage of facies of the Irish Valley at this section is believed to represent various subenvironments of a tidally influenced delta. The large sandstone bodies contain many features characteristic of large sand shoals reported from tidal channels of modern tidally influenced deltas. Large-scale trough and planar crossbedding in the lower parts and the presence of asymmetric (current), symmetric (wave), and linguoidal ripples with variable crest orientation near the upper parts of the sand bodies are evidence for their tidal origin. In addition, the large sandstone bodies interfinger at their bases with Lingula-bearing shale, show numerous internal pause planes of shale, are comprised of well-sorted and quartz-rich grains, and are tabular, not cigar-shaped. The red mudstones containing Lingula are thought to represent brackish-water lagoons and bays. Marsh sedimentation is inferred from red siltstones with abundant root traces. Thin, rippled, quartzitic, fossiliferous sandstones occurring at the tops of many of the sandstones represent marine transgressions, produced during periods when rates of sediment supply were low, a condition produced by lateral shifting of distributaries or variation in discharge, or produced by eustatic sea-level changes. Figure 5 is a diagram of the tidally influenced delta at Port Matilda.

The Sherman Creek Member occupies most of the strike valley to the west of this outcrop. The lack of resistance to erosion and the few small outcrops available lead to the conclusion that the principal lithologies are red mudstone and siltstone, which is in contrast to the greater abundance of sandstone in the corresponding section at Stop 7 in the Milesburg section.

LEAVE STOP 3 and proceed west on Route 322.

0.80 17.20 **Stop 4: Duncannon Member of the Catskill Formation.**
Discussants: E. G. Williams, R. Slingerland, and A. W. Rose.

The rocks exposed at this locality (Fig. 6) are the fining-upward cycles of the Duncannon Member of the Catskill Formation. They resemble those described at Stop 8 of the Milesburg section (Plate 1; compare section 7, 2830-3320 ft to section 5, 2980-3250 ft), but they differ in that they are much thicker and contain relatively more sandstone and less red mudstone and shale. Several smaller cycles, each about 50 feet thick, occur low in the outcrop and are similar to those at Stop 8. Unique to this locality are the much thicker cycles commencing 200 stratigraphic feet above the base of the section. The first cycle is 133 feet in thickness, 126 feet of which is sandstone. Following an erosional base, the sandstone has a 3-foot-thick basal concentration of interformational conglomerate consisting of siltstone and shale clasts, calcareous nodules, carbonaceous wood and plant fragments, and scattered white quartzitic pebbles. Thin conglomeratic units of this composition rest on scoured surfaces at various levels throughout the sandstone body, breaking it up into a number of distinctive morphologic units which give the sandstone body a

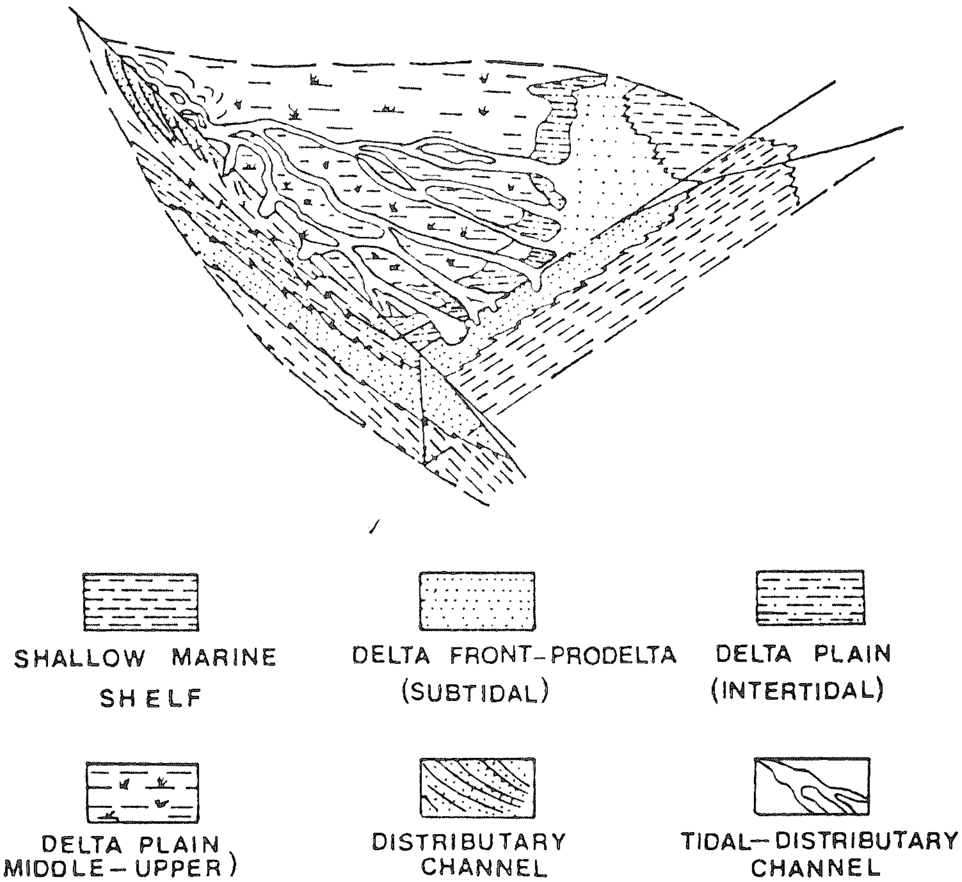


Figure 5. Paleogeographic model illustrating the various subenvironments and facies of the inferred tide-dominated delta at Port Matilda (from Rahmanian, 1979).

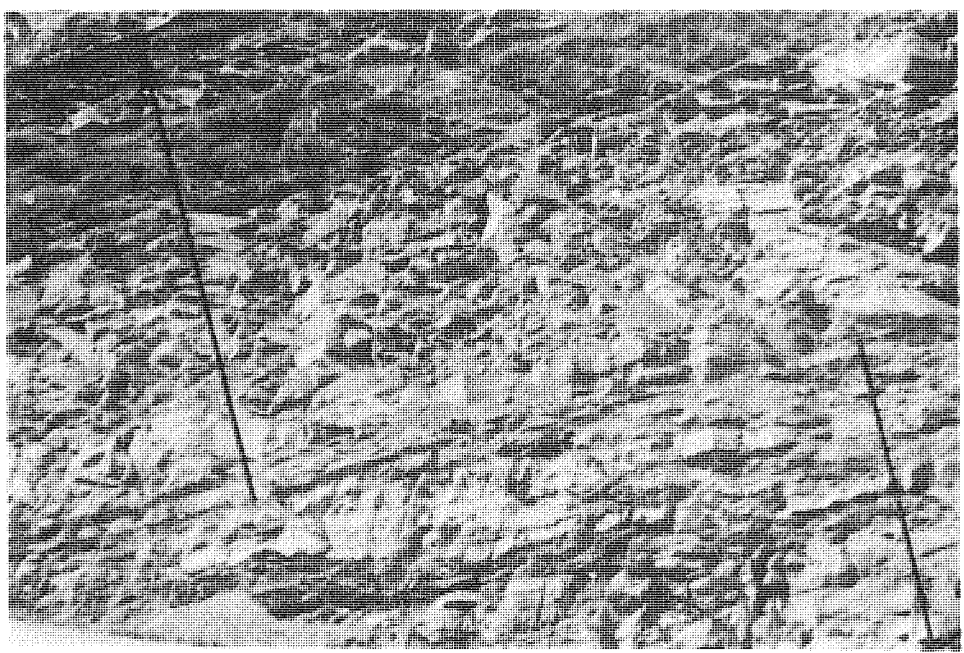


Figure 6. Exposures of the Duncannon Member of the Catskill Formation at Stop 4 containing a 133-foot-thick, fining-upward alluvial cycle denoted by the lines.

composite appearance. Individual sandstone units are green, fine- to medium-grained, micaceous and carbonaceous, and often complexly crossbedded. Measurements of trough and planar crossbedding gives a northwest transport direction (Williams, this volume, Fig. 8). The sandstone body fines upward and eventually grades into a 7-foot interval of red, thin interbeds of siltstone, shale, and claystone, containing many root traces and desiccation cracks. Associated with the rooted mudstone is a 1.5-foot-thick zone of numerous large and small, rounded calcareous nodules, which is interpreted to represent a partially preserved soil horizon. Directly above the soil is 4 feet of green, burrowed, fossiliferous siltstone and silty shale. The fossils are calcareous brachiopods. Stratigraphically above this locality to the west are several other Duncannon cycles of similar thickness and lithology.

The exceptionally thick sand bodies, within which are preserved a large number of lenticular sandstone units, and the scarcity of fine overbank deposits indicate rivers of moderate to low sinuosity whose channels were meandering to anastomosing. In this interpretation, the accretionary bedding so obvious in the outcrop was formed by lateral migration of mid-channel pebbly sand bars as in the Brownstones of southwestern England (Allen, 1983). The occurrence of thin transgressive deposits within the Duncannon would seem to indicate that the area of sedimentation was close to shore and may represent braided distributaries on the upper delta plain.

The origin of the various colors of the Catskill also will be considered at this locality. Facts relevant to this problem are:

1) Composition

Composition	Red siltstone	Green-gray siltstone
Fe_2O_3	3.05%	3.23%
Fe^{3+}/Fe^{2+}	1.68%	0.72%
Grain size (ϕ)	3.84	4.12
Quartz %	24.65%	24.33%
Number of samples	20	19

2) The red color is produced by hematite, the green color by chlorite.

3) Red siltstones, although containing root traces, never have any carbonized remains of plant roots, stems, leaves, or spores, whereas the green and gray siltstones and fine sandstones in the fining-upward cycles generally contain abundant and various types of organic matter.

4) *In situ* calcareous nodules occur in red mudstones at the top of many cycles.

5) The red mudstones of the Irish Valley Member are brackish water in origin and were deposited in bays and lagoons.

6) Black and gray shale chips as well as deformed layers of black shale occur in many of the gray channel sandstones.

We interpret these facts to mean that sediments transported by the braided rivers of the Duncannon were initially gray and that the red mudstones and siltstones at the top of the cycles are overbank and levee deposits which have been oxidized *in situ* in an arid climate. Evidence for the latter interpretation comes from the calcareous nodules which are interpreted to be caliche deposits of a desert soil. The evidence for *in situ* oxidation of green muds to red ones is the occurrence of the black and red shale chips found in the gray channel sandstones.

We reason that these represent overbank muds which slumped into the adjacent rivers during channel migration and bank undercutting. Because the red fragments exhibit no reduction rims, we conclude that the black fragments were black when eroded. The absence of such deposits at the tops of the cycles suggests to us that all the dark sediment has been oxidized to red.

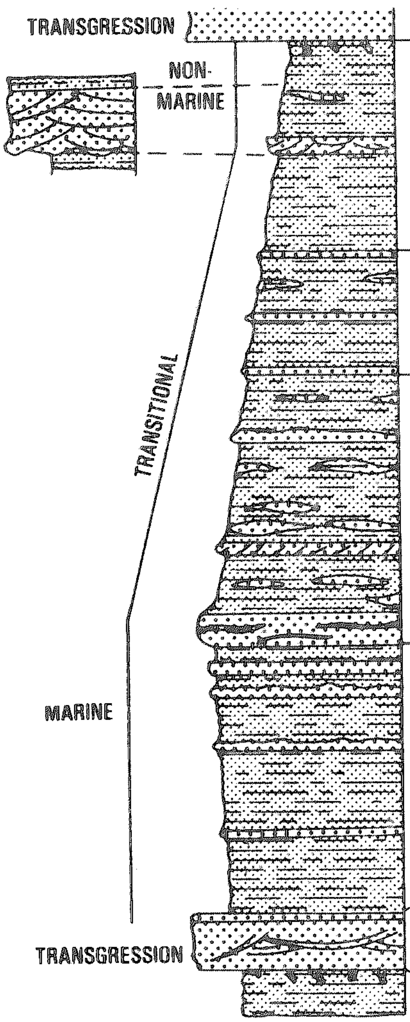
The red mudstones and shales of the Irish Valley Member have a different origin. Because they were deposited in lagoons and bays, it is unlikely that they could have been oxidized in place. We believe that the red mud and silt was transported from the red floodplains of the inactive coastal plains (Sherman Creek Member), where the red overbank sediments were being constantly eroded by meandering streams. Sedimentation was sufficiently rapid in delta-front environments that reduction did not occur even though the bays, with abundant fossils, may have been reducing.

LEAVE STOP 4 and proceed west on Route 322.

0.40	17.60	Rocks exposed on the right are the upper part of the Duncannon Member of the Catskill Formation.
0.40	18.00	Rocks exposed in this and the next large roadcut are sandstones of the Mississippian Pocono Formation. This sand body resembles those of the Duncannon in exhibiting multilateral sandstone bodies bounded by erosion surfaces and exhibiting trough and planar crossbedding. At the base of some channels are thin transported coals. No overbank shales have been observed. In comparison to the underlying sandstones of the Duncannon and to Pocono sandstones at the Milesburg section, the Pocono at this locality is much thicker, is coarser grained, and exhibits little or no overbank deposits. We interpret the Pocono at this locality to be a braided river, the principal river that supplied the delta located in the Port Matilda area.
2.1	20.1	Exposures on both sides of the road consist of the Loyalhanna Formation, a calcareous quartzite, which represents a transgressive shelf sand, and so marks the end of this major Upper Devonian Mississippian regressive sequence.
3.6	23.7	The road is descending along the southeastern dip-slope of the Philipsburg coal basin. The rocks exposed in various roadcuts from here to Black Moshannon State Park are in the Pottsville and Allegheny Groups of Pennsylvanian age.
1.6	25.3	TURN RIGHT directly before Dairy Queen.
0.2	25.5	TURN RIGHT on Route 504 East.
8.6	34.1	LUNCH STOP at Black Moshannon State Park.
		LEAVE LUNCH STOP continuing East on Route 504.

6.7	40.8	Edge of the Allegheny Escarpment with view of Appalachian Mountains to the southeast.
5.2	46.0	TURN LEFT on Route 220 North.
4.0	50.0	TURN LEFT on Route 144 North in town of Wingate.
2.9	52.9	Stop 5. **Irish Valley Member of the Catskill Formation, Runville, Pa.** Discussants: E. G. Williams and R. Slingerland.

The rocks exposed at this locality consist of alternating red, nonmarine sandstones, siltstones, and mudstones and marine gray and green sandstones and shales (Plate 1, section 6). These sequences are repeated several times and are very similar to those described by Walker and Harms (1971) in the Catskill of the Susquehanna Valley area, where they were called "motifs" to avoid the more rigid concept of cyclothem. They are characteristic of the Irish Valley Member of the Catskill Formation at this and other sections south of Port Matilda. Regionally, two types are recognized--a silty-muddy motif and a sandy-silty motif, but only the former occurs at Runville and adjacent localities. The lithology and environmental interpretations of a typical silty-muddy motif from this area are illustrated in Figure 7. Five subfacies can be recognized in these motifs, namely: 1) basal bioturbated sandstone, 2) green shale and silty shale, 3) green sandstone-mudstone, 4) red laminated siltstone, and 5) red massive mudstone and siltstone. The cycles vary in thickness from 5 to 65 feet and average about 15 feet. Each motif has an almost planar and sharp basal contact which separates the basal green marine sediments of the motif from the red mudstone of the underlying motif. Evidence suggests that each motif started with an initial transgressive event, now represented by the bioturbated basal sandstone and the overlying green fossiliferous silty shale facies. The presence of brachiopods and crinoidal debris and burrows of marine affinity implies a normal marine environment for this facies. Absence of thick and winnowed sand bodies in the transgressive phase reflects a rapid rate of transgression coupled with a high rate of mud and silt supply. In contrast, the sandy-silty motifs (best developed at Entriken and south) contain a basal quartzitic sandstone facies up to 4 m thick comprised of large solitary sets of bipolar, planar cross-strata, indicating reworking by tidal currents. The transgression was followed by a stage of very slow sedimentation through which extensive biological reworking of the deposited sediments, including the transgressive beaches, took place. Finally, marine deposition of mud gradually changed to deposition of thin, well-sorted sandstone with symmetric ripples (probably wave generated) as the sea shoaled. Above the shoaling phase occurs a fining-upward, regressive, tidal-flat sequence. The complexly bedded, thin quartzitic sandstone and the interbedded burrowed and flaser-bedded green siltstone and shales are interpreted to represent low- to mid-tidal-flat facies. The overlying red laminated siltstone and red massive mudstone are thought to occur in the high to supratidal range. The presence of root marks, desiccation and mud cracks, and carbonate nodules supports upper to supratidal-flat environments. The sandstone units which cut through the high-tide mudflat sediments may represent meandering tidal channels which drain the upper tidal flat and supratidal environments. However, the absence of such sand bodies in the underlying mid and lower tidal-flat deposits requires another interpretation, namely that they might represent small meandering streams which drained part of the coastal area. The scarcity, thinness, and fine grain size of these sandstone units suggest that only small and sluggish local streams drained this part of the coastal-plain area. This indirectly suggests that a coastal-fringe area, in the



Facies	Sedimentary structures	Interpretation
Red massive mudstone	Massive; desiccation cracks; abundant root traces; occasional interbedded red, lenticular, fining-upward sandstone bodies	High-tide to supratidal mudflat
Red laminated siltstone	Desiccation cracks; horizontal (rhythmic), and wavy siltstone beds	
Green sandstone-mudstone facies (dominantly mudstone)	Flaser, wavy and lenticular bedding; blocky and massive in places; small-scale solitary, interference, and linear, clay-draped ripples; burrowing and bioturbation common	Low-mid tidal flat
Green fissile shale and silty shale (fossiliferous)	– Silty shales: Finely laminated – Sandstones: Rippled to micro-crossbedded – Bioturbated intervals common	Low-energy shallow marine shelf-shoreface
Basal, bioturbated sandstone	– Extensive bioturbation – Horizontal to low-angle crossbedded	– Transgressive sand (beach)

Figure 7. Idealized vertical facies sequence of the silty-muddy motif of the Irish Valley Member derived from pooled data from sections north and south of Port Matilda (modified from Rahmanian, 1979).

region of the development of the Irish Valley motifs, was flanked by an inactive alluvial plain.

The varying thickness and limited lateral extent of individual motifs leads to the conclusion that they were produced by variation in sediment supply, which resulted from lateral shifts in the position of the deltaic lobe or its distributaries located to the south near Port Matilda. Transgression occurred when the delta was diverted from a particular part of the shoreline, coupled with regional subsidence and compaction. Regression would occur when the delta or its distributaries shifted back to a closer position. Under this condition, increase in flux of sediments by longshore drift would have caused coastline progradation by depositional regression and seaward building of tidal flats.

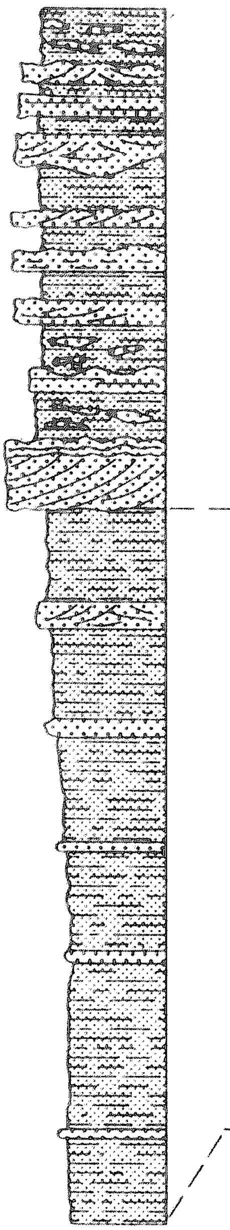
LEAVE STOP 5; TURN AROUND and retrace path to Route 220.

3.0 55.9 TURN LEFT on Route 220 North.

3.0 58.9 TURN LEFT on entrance ramp for I-80 West towards DuBois.

2.1 61.0 Stop 6: Upper part of the Trimmers Rock Formation. Discussants: E. G. Williams and R. Slingerland.

The rocks exposed here are the silty shale and sandstone-shale facies of the upper part of the Trimmers Rock Formation (Plate 1, section 5, 0-400 ft). The lithologic characteristics and the environments are summarized in Figure 8. Sandstones are frequently hummocky cross-stratified with ripples on the upper surfaces, locally



Facies	*Sedimentary structures*	*Interpretation*
Sandstone-shale facies Light- to olive gray interbeds of fossiliferous sandstone and silty shale	- Sandstones: Large-scale and solitary to small-scale crossbedding; linear (symmetric and asymmetric) and interference ripples; horizontal wavy (hummocky) and flaser bedding; beds 5-20 ft thick; quartz-pebble bases; many with 4-ft-thick solitary planar cross-strata dipping northeast - Silty shale: Horizontal, wavy bedding - Bioturbated horizons common	Shallow marine subtidal shelf (sand waves and ridges)
Silty shale facies Gray olive-gray, fossiliferous silty shale and siltstone with thin sandstone layers interspersed	- Siltstones: Horizontal, wavy, and/or ripple lamination - Sandstones: Small- to micro-scale crossbedding; linear and interference ripples; wavy and horizontal lamination (hummocky cross stratification) - Bioturbated horizons common	Shallow marine shelf

Figure 8. Idealized vertical facies sequence of the uppermost 250 to 500 feet of the Trimmers Rock Formation derived from pooled data from north and south of Port Matilda (modified from Rahmanian, 1979).

contain quartz pebbles, and have groove and flute casts along bottom contacts. Gray siltstone and shale beds are extensively bioturbated; burrows are usually solitary, oriented both parallel and perpendicular to bedding. Marine fossils occur throughout the interval. Brachiopods dominate but there are also pelecypods, crinoid columnals, bryozoan fragments and encrustations, fish plates, and plant fragments. Shells occur as solitary forms in shales and as coquina layers at the top of many sandstone beds. At most localities, the silty shale facies grades upward into a sandstone-shale facies which is illustrated in the upper half of Figure 8. Grain size, thickness, and the number of sandstone beds increase.

These facies are interpreted to have formed within a shallow marine shelf environment. The presence of hummocky cross-stratification and symmetric ripples in conjunction with small-scale crossbedding suggests a position above wave base in the shelf environment, where both waves and currents operate. Small-scale grain size variation found in the siltstone and silty shales indicates fluctuation in sediment supply and current velocity, which could be due to variation in tide- or storm-generated currents or both. Petrographically the sandstone beds resemble those of the alluvial sandstones of the Sherman Creek Member in terms of grain size and mineral composition.

LEAVE STOP 6 and proceed west on I-80.

1.4 62.4 Stop 7: Fining-upward alluvial cycles of the Sherman Creek Member. Discussants: E. G. Williams and R. Slingerland.

Rocks exposed at this locality (Fig. 9) are the fining-upward alluvial cycles of the Sherman Creek Member, consisting of interbeds of thick red mudstone, siltstone, and to a lesser degree red and green, fine- to very fine grained sandstone (Plate 1, section 5, 1720-2230 ft). The cycles start with dominantly red, lenticular to sigmoidally bedded sandstone units which usually possess basal erosional contacts and gradational tops. The basal sandstone units grade upward to a usually thick sequence of interbedded red siltstone with dominantly red mudstone at the top of the cycle. Cycles vary in thickness from 5 to more than 30 feet and average 20 feet. Small and large bone fragments and fish dermal plates occur together with basal interformational shale clasts. Crossbeds dip to the northwest (Williams, this volume, Fig. 8). The siltstone and mudstone units are usually massive with blocky and crumbly weathering, and contain small root traces and small calcareous nodules in the uppermost part of the cycles.

The inferred environments for the Sherman Creek cycles are those associated with shallow meandering rivers. The lower sandstone represents point-bar deposits; the laminated fine sandstone and siltstone units are interpreted as levee deposits. The red mudstones represent floodplain deposits as evidenced by desiccation cracks and calcareous concretions. Thin, lenticular sandstones interbedded with the red mudstones are thought to be crevasse deposits. The relative thinness of the basal sandstones and the small width of abandoned channels (8-40 feet) suggest that the rivers were relatively small. A few beds of green quartzitic sandstone and shale occur at irregular intervals throughout the Sherman Creek. These are similar to the marine part of the Irish Valley and are believed to have been deposited in a tidal-flat environment during a transgression.

The general paleogeographic setting for the fining-upward cycles of the Sherman Creek is that of the distal parts of a broad and inactive coastal-plain environment, bordered in the seaward direction by extensive tidal flats (Irish Valley motifs).

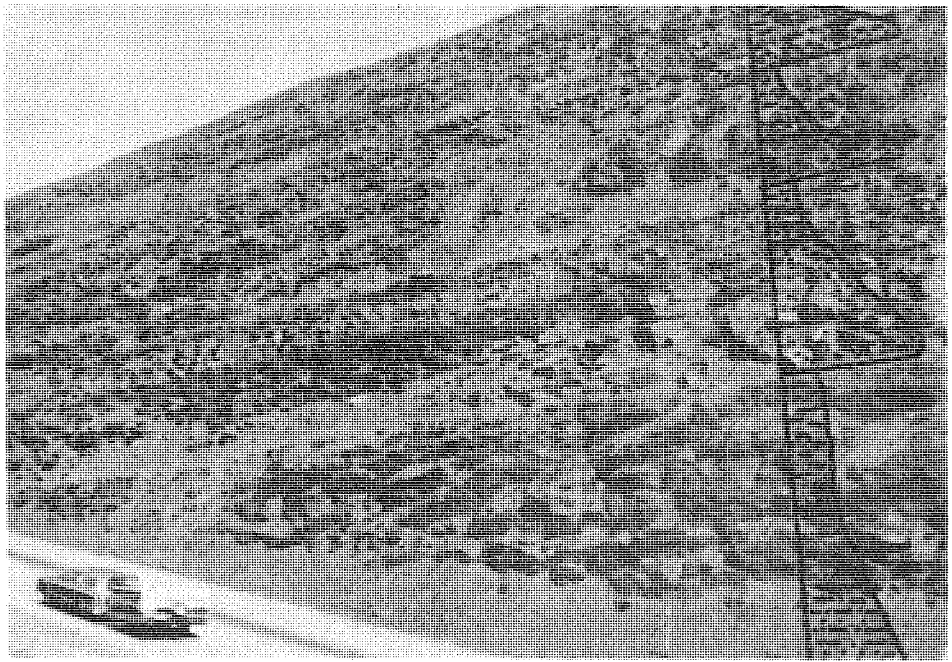


Figure 9. Exposures of the Sherman Creek Member of the Catskill Formation at Stop 7. Note the fining-upward cycles and the predominance of mudstone and siltstone.

The coastal plain had a very low gradient as indicated by the several marine transgressions within the alluvial sequence.

LEAVE STOP 7 and proceed west on I-80.

2.3 64.7 Stop 8: Fining-upward cycles of the Duncannon Member of the Catskill Formation. Discussants: E. G. Williams and R. Slingerland.

Rocks exposed at this locality (Fig. 10) are the fining-upward cycles of the Duncannon Member of the Catskill Formation. The cycles, which average about 50 feet in thickness, start with a sandstone unit which makes up approximately 60 percent of

Figure 10. Exposures of the Duncannon Member of the Catskill Formation at Stop 8 containing two fining-upward alluvial cycles.

the cycle, has an erosional base and gradational top, and grades upward into layers of finer sandstone, siltstone, and shale. Concentrations of drifted plant debris, shale pebbles, and calcareous nodules are usually found at the base. The lower half of the sandstone is usually green gray in color, grading upward to red as it becomes finer. The basal green sandstone is medium grained, poorly sorted, usually micaceous and carbonaceous, and thick to massively bedded at the base, and commonly shows an upward decrease in bed thickness. The red sandstone is fine grained and silty; it grades upward into red siltstones and shales of the upper parts of the cycle. Large-scale trough and planar crossbeds occur in the lower parts of the basal sandstone and exhibit northwestward dip directions. The red siltstones and mudstones commonly contain desiccation cracks and abundant calcareous nodules at the top of the cycle. These fine sediments make up less than 40 percent of the cycle.

In comparison to the underlying Sherman Creek cycles, those of the Duncannon contain thicker, coarser grained sandstones and thinner red overbank siltstones and mudstones, suggesting larger channels of lower sinuosity, as discussed at Stop 4.

LEAVE OUTCROP and proceed west on I-80.

1.8	66.5	Outcrops on right are the lower part of the Pocono Formation. All outcrops from this locality to the Snow Shoe interchange are in the Pocono Formation.
3.0	69.5	EXIT Route I-80, turn left at end of exit ramp, cross bridge, turn left and reenter I-80 heading east.
10.6	80.1	EXIT I-80 on Route 220 South towards Milesburg.
1.2	81.3	EXIT RIGHT on Route 150 South towards Milesburg.
3.4	84.7	STOP LIGHT in the center of Bellefonte, Pa. 100 m on the right is the Big Spring from which Bellefonte gets its water; for further information, see Parizek and White (this volume).
6.6	91.3	STOP LIGHT at junction of Routes 150 and 64. Continue south (straight) on Route 64 towards State College.
4.2	95.5	STOP LIGHT at corner of College Ave. and Atherton St. in State College; TURN LEFT on Atherton St.
1.5	97.0	Holiday Inn on right.

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FIELD TRIP #3

APPLICATION OF QUATERNARY AND TERTIARY GEOLOGICAL FACTORS TO ENVIRONMENTAL PROBLEMS IN CENTRAL PENNSYLVANIA

by
Richard R. Parizek and William B. White

Introduction

Nittany and adjacent valleys near the western edge of the Valley and Ridge province are challenging but informative areas in which to study the evolution of drainages and erosion surfaces in the central Appalachians because rocks ranging in age from Late Cambrian to Pennsylvanian are exposed within less than 10 miles of each other, and the structural and topographic style changes from Valley and Ridge to the Allegheny Plateau setting.

Nittany Valley is the most distant valley from the Atlantic Ocean within the Ridge and Valley of Pennsylvania. Stream terraces and erosional surface remnants are more likely to be preserved in such an interior valley because more time is required for drainage networks to adjust headward in response to changes in sea level and tectonic uplift. Also, there is greater relief between stream valleys and uplands, and terrace remnants are likely to have greater vertical separation making them easier to identify when compared to valleys near the eastern margin of the province.

The northern edge of the province was overridden by continental glaciers of Kansan(?), Nebraskan, Illinoian and Wisconsinan age. Their deposits blanket older erosional surfaces and soil deposits that pre-date Pleistocene glaciation. The region was subjected to repeated climatic changes associated with the advance and retreat of continental glaciers that helped produce an array of periglacial deposits, landform, and erosional surfaces of variable age. These latter deposits and landforms, although posing challenges of their own, represent only minor alterations in the more regional landscape. These youngest deposits, together with erosional and constructional landforms are easier to date hence, help to set minimum ages for erosion surfaces and help estimate rates of erosion.

On the field trip, we bring up some of the old questions again. Given present knowledge of soils, landforms, drainage patterns, and their evolution, what can be said about the developmental history and time scale of development of the Nittany Valley.

Sevon's (1985) brief summary of previous work on Appalachian drainage covers the main concepts involved in the discussion of drainage and landscape development in Pennsylvania. He lists items important to these concepts which require further discussion. These include:

(1) The position of the drainage divide after the Alleghenian orogeny.
(2) The direction of initial drainage after this orogeny.
(3) The time of origin of southeastern drainage flow.
(4) The reality of peneplaination in Pennsylvania.
(5) The reality of Cretaceous transgression onto a peneplained surface.
(6) The reality of repeated uplift.
(7) The relation of transverse drainage to lithologic or structural weakness.
(8) The rate of denudation in the Appalachians and the age of the landscape.

Geologic Setting

Geology

Nittany, Penns and adjacent karst valleys are underlain by 6,000 - 8,000 feet of interbedded limestones and dolomites of Cambrian and Ordovician age (Figures 1 and 2) which were initially capped by 3200 feet of shales and ridge forming sandstones and many thousands of feet of younger strata Devonian through Pennsylvanian in age. These have been folded into gently plunging anticlines and synclines, and overturned to steeply dipping asymmetric folds displaying structural relief of 100 to several thousand feet, cut by numerous normal and thrust faults, weathered and eroded under climatic conditions varying from arid, tropical to subtropical to periglacial during the Pleistocene Epoch, to temperate-rainy at present where the mean annual precipitation approaches 38-43 inches. Elements of the present landsurface may reflect chemical weathering from Late Cretaceous to the present.

Rocks most resistant to erosion form ridge crests which surround canoe-shaped lowlands underlain by less resistant carbonate rocks and shale. Ridges surrounding Nittany, Penns and Kishacoquillas Valleys normally are double. The innermost ridge is underlain by the Bald Eagle or Oswego Sandstone, the higher and outer most ridge by the Tuscarora Quartzite (Stop 1). A saddle frequently is developed on the Juniata Formation which is largely composed of siltstone and shale. It lies between the Oswego and Tuscarora. Water gaps commonly are developed in the Oswego Sandstone or in the inner ridge and major drainages tend to cut through both ridges (Parizek et al., 1971). Minor sags and wind gaps reflecting narrow zones of weakness are repetitous in both ridges.

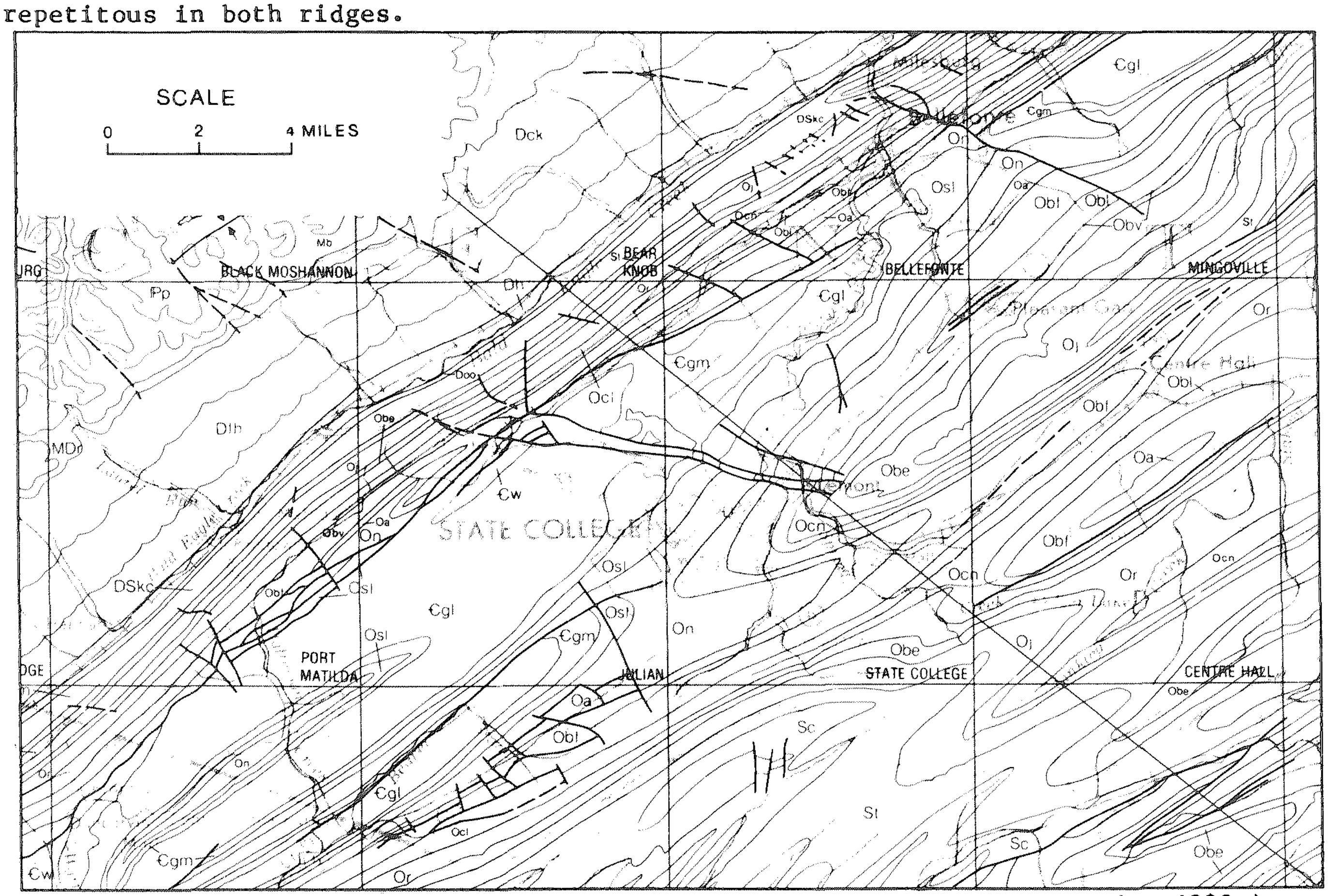


Figure 1. Geologic map of the Nittany Valley area (geology from Berg et al., 1980.)

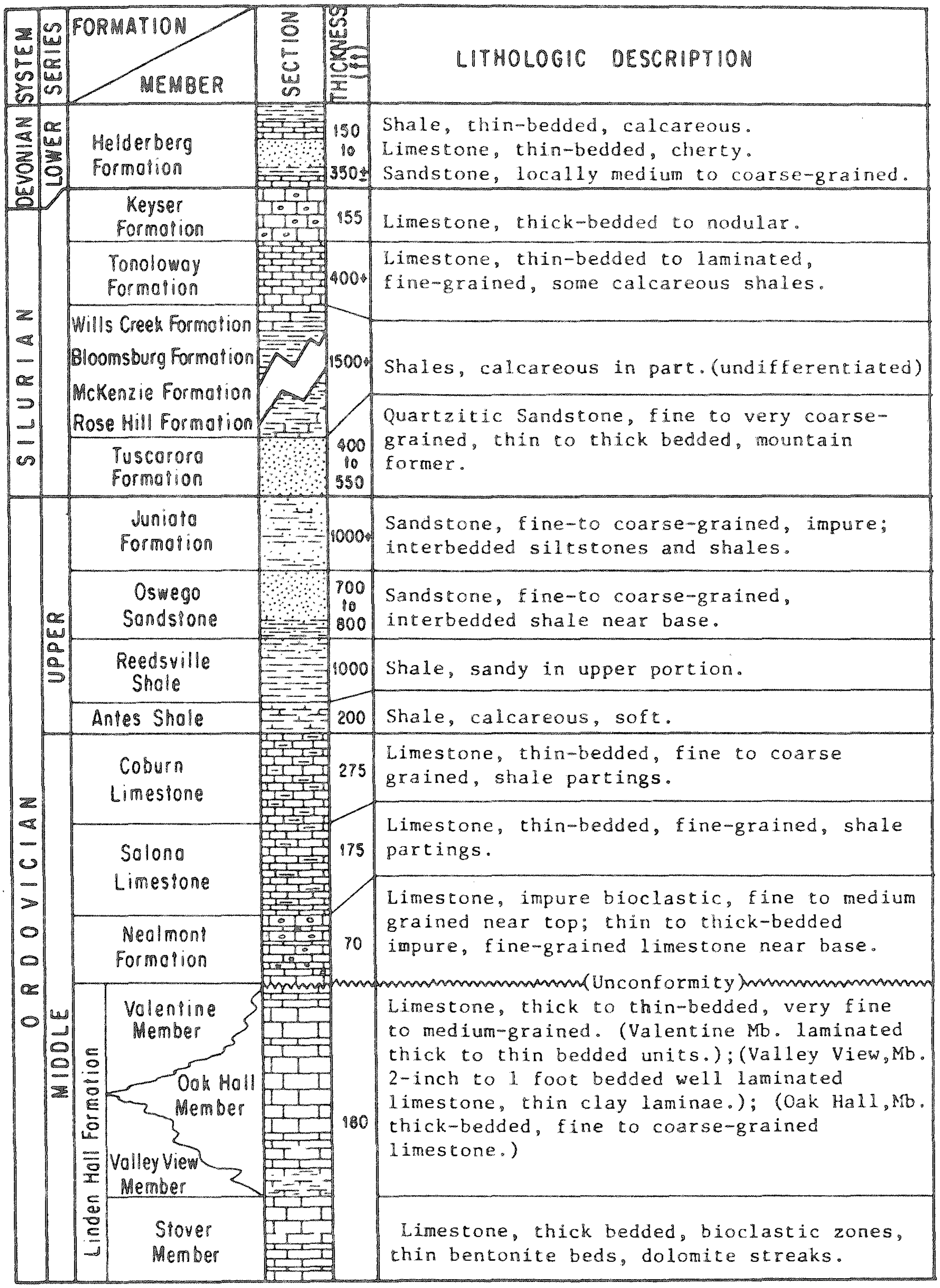


Figure 2. Stratigraphic section for Nittany Valley (from Parizek et al., 1971).

SYSTEM	SERIES	FORMATION / MEMBER	THICKNESS (ft)	LITHOLOGIC DESCRIPTION
ORDOVICIAN	MIDDLE	Snyder Formation	190	Limestone, 4 inch to 1-foot beds, fine to medium grained; interbedded dolomite, oölitic beds, mud-cracked beds, clay partings, and coarse bioclastic beds.
ORDOVICIAN	MIDDLE	Hatter Formation	100	Limestone, 4-inch to 2-foot beds, fine to medium grained, with laminated argillaceous and arenaceous dolomite, fossiliferous, worm borings. (Unconformity)
ORDOVICIAN	MIDDLE	Clover Limestone	80	Limestone, 2-inch to 2-foot beds, fine to very fine grained, laminated with fine to coarse-grained limestone.
ORDOVICIAN	MIDDLE	Milroy Limestone	300±	Limestone, fine-grained, silty; laminated with wavy dolomitic bands.
ORDOVICIAN	LOWER	Bellefonte Dolomite: Tea Creek Member	200	Dolomite; fine-grained to sublithographic, thin shale partings, gashed weathered surfaces; 1 to 4 foot-beds.
ORDOVICIAN	LOWER	Bellefonte Dolomite: Dale Summit Member	0 to 14	Sandstone, fine to coarse-grained, conglomeratic.
ORDOVICIAN	LOWER	Bellefonte Dolomite: Coffee Run Member	1000	Dolomite, interbedded fine-to medium-grained, cyclic successions.
ORDOVICIAN	LOWER	Axemann Limestone	400	Limestone, fine to coarse-grained, oölitic, interbedded thin layersof impure dolomite, fine to medium-grained, partly conglomeratic, chert locally.
ORDOVICIAN	LOWER	Nittany Dolomite	1200	Dolomite, fine-to coarse-grained alternating in cyclic manner, spherical chert nodules, oölitic chert, thin limestone and sandy beds.
ORDOVICIAN	LOWER	Stonehenge Limestone	600	Limestone aphanitic to fine-grained, argillaceous and dolomitic in part, flat pebble conglomerate abundant.
CAMBRIAN	UPPER	Gatesburg Formation: Mines Member	150 to 230	Dolomite, interbedded coarse-to fine-grained, chert abundant, oolitic, thin sandy beds near base, vugular.
CAMBRIAN	UPPER	Gatesburg Formation: Upper Sandy Member	500±	Dolomite; with interbedded orthoquartzites, and sandy dolomites, some shaly dolomites, fine to medium grained, vugular.
CAMBRIAN	UPPER	Gatesburg Formation: Ore Hill Member	260	Dolomite; fine to medium grained thick-bedded near top, fine-grained argillaceous dolomite near center; coarse-grained, massive bedded at base.
CAMBRIAN	UPPER	Gatesburg Formation: Lower Sandy Member	700±	Dolomite, interbedded orthoquartzites, thin-bedded fine-to medium grained dolomite, medium to coarse-grained orthoquartzite; thin bedded shaly dolomite.
CAMBRIAN	UPPER	Warrior Formation	600	Limestone, in part dolomitic, thick-bedded, with thin-bedded shale and sandy units.

Figure 2. (continued)

Prominent Landscape Elements

The mountain ridges surrounding Nittany Valley display a nearly accordant elevation ranging from 2000 to 2400 feet which is lower than the sandstone capped eastern margin of the Appalachian Plateau located to the west (Stop 1). This surface was referred to as the Schooley Peneplain by Davis (1899). The relatively flat inner valley upland ranges from 800 to 1200 feet in elevation. It is underlain by carbonate rocks and some shales and is covered by residual soils (sapolite) of variable character and thickness (Figure 3). This surface was referred to as the Harrisburg Peneplain by Davis.

Spring Creek, Buffalo Run, Logan Branch, Slab Cabin Run, Nittany Creek, and Little Fishing Creek are the principal drainages for the northeastern portion of Nittany Valley. Bald Eagle Creek is the local base level for these streams which in turn, drains into the West Branch of the Susquehanna River near Lock Haven.

Halfmoon Creek, Spruce Creek, and Warriors Mark Run drain the southern portion of the valley via the Little Juniata River which joins the Susquehanna River at Amity Hall. The Juniata River cuts through Nittany Valley along one of the state's longest lineaments, the Mt. Union-Tyrone lineament which is the subject of a companion trip. The Juniata serves as a southern base level of erosion to Nittany Valley. Groundwater levels are controlled by elevations of the Juniata and by sandstone and shale beds truncated by the Juniata where it crosses Tussey Mountain.

These streams are incised into the valley upland to depths of 50 to 300 or more feet and often display steep cliffs. Valleys vary in width to just accomodate the width of tributary streams or they may be wide enough to support meanders and well developed alluvium. The floors of these valleys have been correlated with the Sommerville erosion surface by Davis and were believed to have been incised in response to regional uplift.

Floodplains display variable widths along these valleys having characteristics of both youth and maturity often only a short distance down the same valley. Changes in valley width occur where these rivers and creeks traverse bedrock units of varying resistance. Streams in narrow valley sections often are influent whereas they are often effluent along wider sections of the same valley. Many of the smaller tributaries to these streams are karst underdrained valleys that show varying stages of karst development. The fluvial origin of these underdrained valleys is obvious even where they lack floodplain deposits due to internal erosion and where their original fluvial profile has been disrupted by underlying karst processes.

The most prominent floodplains occur above shale and carbonate bedrock upvalley from more resistant bedrock that served as a temporary base level of erosion. Examples include the confluence of Spring and Slab Cabin Creek at Houserville (Stop 5), the flat at the 28th Division Shrine at Boalsburg upstream from where Spring Creek cuts across Nittany Mountain, a flat just above the Borough of State College (Water) Authority well field developed on Slab Cabin Run below Shingletown and elsewhere.

The time of incision of these valleys is hard to date but they existed in time to allow the development of at least one prominent stream terrace located just above the modern floodplain that predates the Sangamonian and alluvial-colluvial fan complexes of pre-Sangamonian and most likely, Woodfordian age.

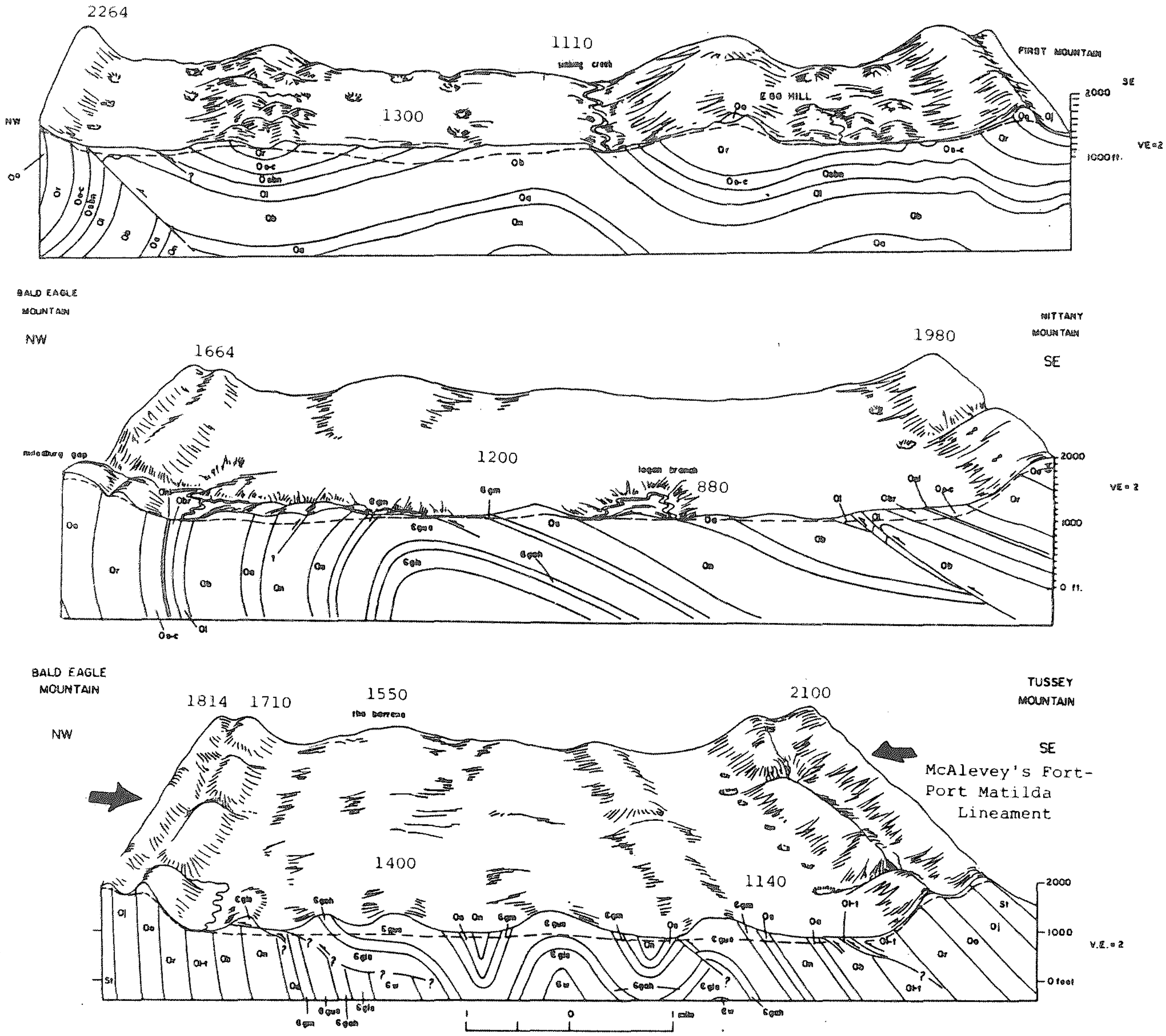


Figure 3. Geological cross-sections of Nittany Valley.

Karst Landforms

The carbonate rock-floored valleys of the Valley and Ridge Province are sculpted into a somewhat subdued karst topography. The three principal landforms are swallow holes and minor blind valleys, closed depressions, and caves. To these should be added a cutter and pinnacle topography etched onto the bedrock of certain formations but generally mantled by thick soils and not apparent unless the soil is removed. The cutter and pinnacle topography is, however, of great concern to those designing foundations for buildings, holding ponds, and other structures to be erected on the karst.

Although the solubilities of calcite and dolomite are nearly equal, the kinetics of calcite dissolution is much more rapid than the kinetics of dolomite dissolution. As a result, karst features are developed to a greater degree on the limestones than on the dolomites. Most of the limestone lithologically suited for extensive karst development occurs in the Champlainian group near the top of the carbonate section. Because of the anticlinal valleys, the outcrop pattern generally has dolomite making up most of the center of the valley with the cavernous limestones occurring as parallel bands at the valley margins. As a result, closed depressions and caves are mainly located in these bands parallel with the mountain flanks. Sinkholes are both more sparse and more subdued in the dolomite portions of the valley and except on close inspection, Nittany and the other valleys do not look like karst to the casual observer.

Small surface streams arise in catchment basins of a few square miles on the flanks of the mountain ridges and flow out onto the carbonate valleys. With few exceptions, these allogenic streams sink at the contact with the Champlainian limestones later to resurge at the large karst springs. Sometimes the water merely disappears into its bed with a dry channel maintained down stream to be used in flood flow. Others have cut minor transverse valleys into the valley uplands and sink at the blind footwall of the incised valley. The depth of incision varies from a few feet to on the order of 100 feet. In some blind valleys streams flow into open cave entranes but mostly the water is simply lost in silt, collapse breccia, and debris. The allogenic mountain streams are an important source of recharge to the carbonate aquifer system accounting for more than 60 percent of recharge and only 20 percent of the area.

Closed depressions occur in a range of sizes from shallow swales a few feet deep to sinkholes many tens of feet in depth. Some are of solutional origin, a few represent cave roof collapses, and some are due to soil piping. The latter tend to be ephemeral features. There may be an abrupt collapse of the soil cover during spring thaws followed by the gradual slumping and infilling of the sink. In the more urbanized parts of the regions, soil piping sinks are a land use hazard and their development is exerbated by parking lot runoff, leaking water and sewer mains, and general disruption of overland flow (Parizek et. al., 1971; White et al., 1984). The permanent sinkholes are usually soil-mantled with little bedrock in evidence so that it is not easy to distinguish solution from collapse features. The largest of the closed depression features is Phantom Lake, nearly a mile in diameter but less than 100 feet deep. It is located two miles east of Pleasant Gap and is just off the field trip route. Water floods the sink by rising through sinkholes on the floor of the depression during periods of unusually high precipitation, thus the name. Although surface expression of closed depressions are sparse in most of the Valley, there is little surface drainage and the water falling onto the limestone surface infiltrates through the soil into the solution cavities beneath.

Determining Background

Prolonged sedimentation within the Appalachian geosyncline resulted in the accumulation of a thick sequence of rocks of variable resistance. These were deposited in marine and freshwater environments that reflect a major and prolonged marine transgression starting in the Cambrian and ending by Pennsylvanian and possibly Early Permian time with the deposition of largely freshwater, continental and nearshore marine deposits of Pennsylvanian age. More than 8,000 feet of limestone and dolomite were deposited in shallow Cambrian and Ordovician seas before influxes of clay, sand and conglomerate began to dominate sedimentation by Late Ordovician and Silurian time (Figure 2). With the exception of the accumulation of some limestone units in the Silurian and Devonian, Silurian through Pennsylvanian sedimentation was largely dominated by clastic rocks of variable resistance. These sedimentary rocks were folded and faulted during the Appalachian Orogeny in Permian

Time approximately 250 million years ago. The Nittany Anticlinorium arches from the eastern edge of the Appalachian Plateau nearly ten miles to the southeast. The Warrior Limestone is raised within this arch and represents the oldest rocks exposed in Nittany Valley. Numerous, small anticlines and synclines formed within this anticlinorium which have been subjected to more or less continuous and prolonged erosion to the present.

Ideas on the tectonic style of the Appalachians have changed since the first Pennsylvania Field Conference as new deep drilling data became available and the structural complexity more completely worked out. The mobile oil and gas test well drilled in Nittany Valley near Jacksonville (northeast of Bellefonte) together with the deep tests provided useful insight that lead to the "thinned skinned" tectonic model proposed by Gwinn (1964) and others (Figure 4). This tectonic setting together with the sequence of rocks deposited in the Appalachian geosyncline are important to the understanding of the erosional history of the Appalachians.

Global Plate Tectonic models also are new and shed light on the driving forces that are responsible for regional uplift and stress distribution that produced thrust and tear faults, zones of fracture concentration, and other planes of weakness that assisted weathering and erosion.

A well kept record of these erosional events is hidden in the many thousands of feet of sediment located in the Coastal Plain and continental shelf along with their elusive petroleum resource. Other clues to early history may be contained within the Triassic basin sedimentary record. However, lacking a well defined early tectonic and climatic history for the Appalachians, a detailed account of the early erosional history of this region may never be satisfactorily worked out.

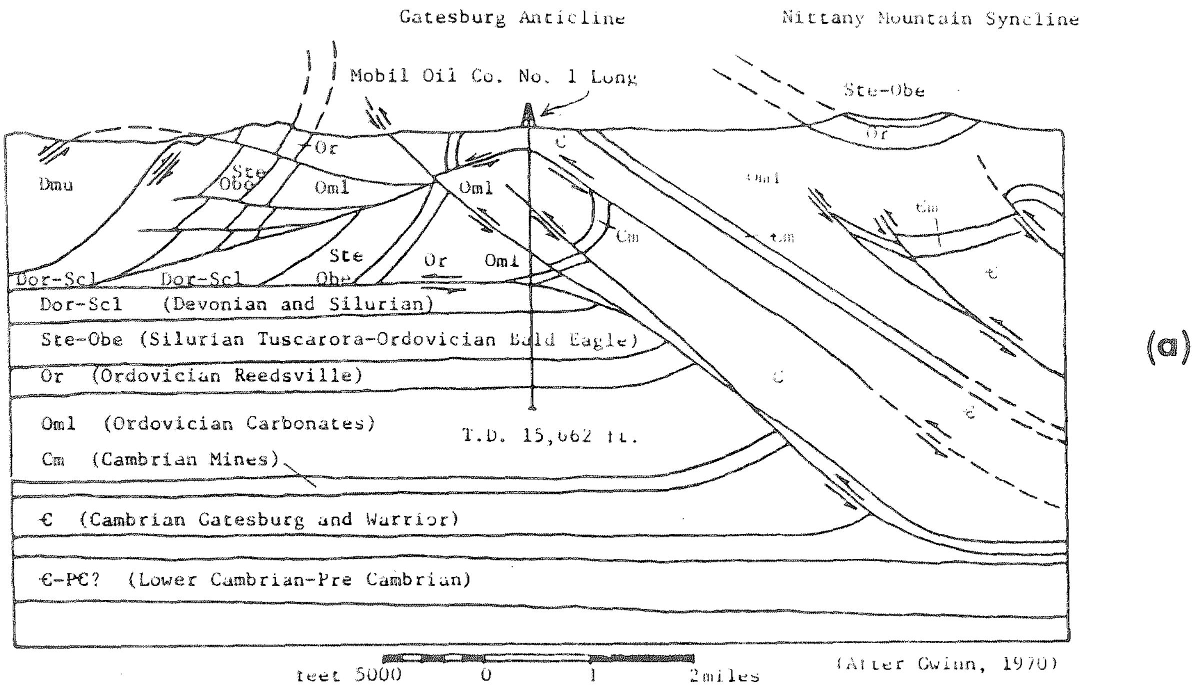
Drainage

The Valley Upland Surface

The most pronounced feature of the Nittany Valley is the upland surface. In the interfluve areas between present day incised drainage, the valley uplands have very low relief. Near Centre Hall on the Penns Creek/Spring Creek divide, on the Penns Creek/Elk Creek divide in Brush Valley, and on the Spring Creek/Spruce Creek divide just west of Pine Grove Mills, the upland surface is nearly flat and truncates geologic structure. Remnants of the uplands can be seen in hill summits and upland flats in other parts of the valley. The uplands do have a measurable slope. In the Penns Valley area to the east, the upland flats are at elevations of nearly 1300 feet. In the area of State College, the uplands are at 1200 feet. Farther to the West, in the uplands bordering the Juniata River gorge the upland flats lie between 1000 and 1100 feet. In a certain vague sense, at least, the old valley floor surface (Harrisburg peneplain) slopes to the present location of the Juniata on the Tyrone lineament suggesting that the lineament may represent the location of very old east-trending drainage.

Incised Drainage

The conceptual difficulty with an old erosion surface dissected by younger drainage is that three out of the four major streams exit Nittany Valley though deep water gaps cut through quartzite in the bounding mountain ridges. If the paleodrainage required only a single exit along the Tyrone lineament, how did these other drainage lines initiate and what determines their location? These questions have perplexed Appalachian geomorphologists since the days of Henry Rogers.



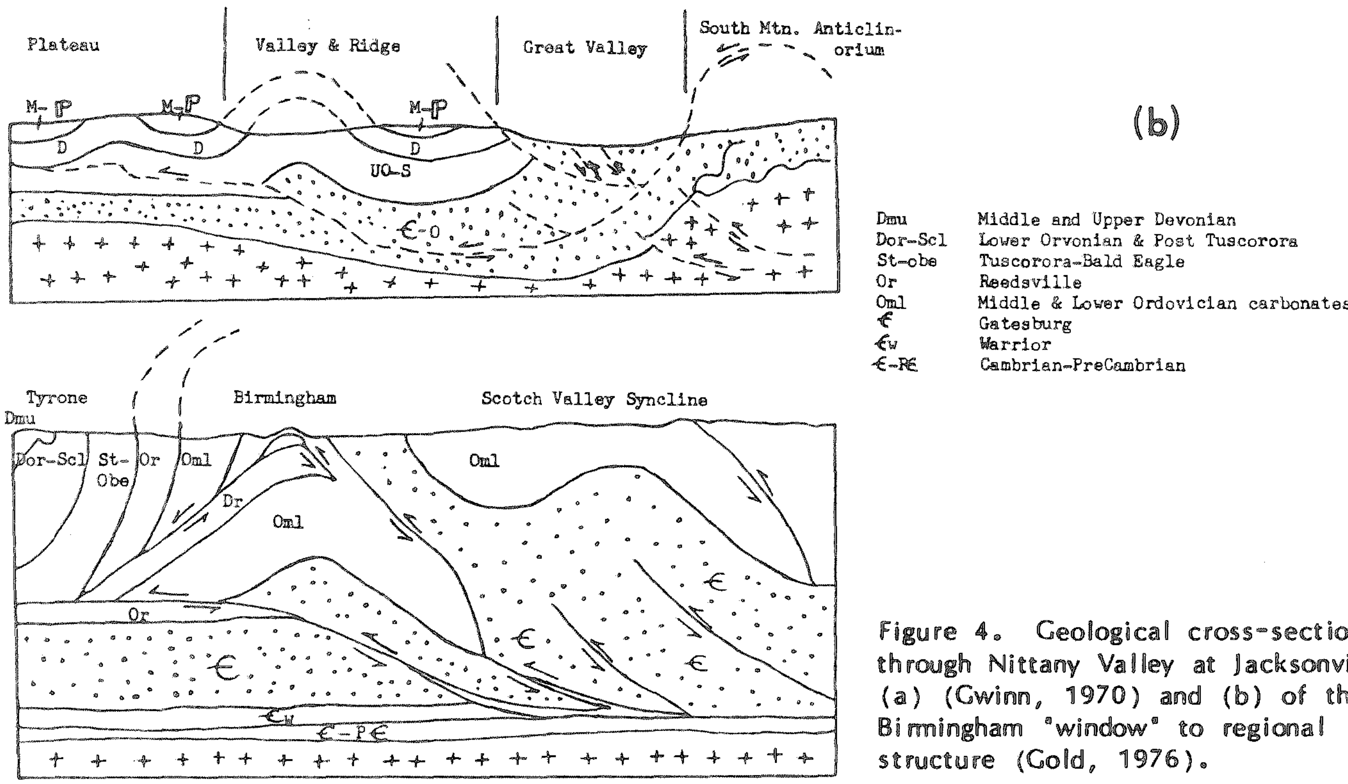


Figure 4. Geological cross-sections through Nittany Valley at Jacksonville: (a) (Gwinn, 1970) and (b) of the Birmingham "window" to regional structure (Gold, 1976).

Lattman and Parizek (1964) noted the rather repetitious spacing of smaller sags, wind and water gaps in ridge tops which they attributed to nearly vertical zones of fracture concentration revealed by fracture traces, lineaments, and minor reverse faults. It was possible to align some wind and watergaps with sinkholes, swallow holes, surface sags and depressions, springs, karst underdrained valleys oriented transverse to stratigraphic and structural strike, straight segments of cave passages, and transverse valleys with permanent streams. Such nearly vertical structures are required to fix the position of drainages in time and space as the landsurface was eroded. The number of such alignments of drainage and the size of associated lineaments far exceeded the number of linear features known or inferred prior to the launching of LANDSAT (Gold, Parizek and Alexander, 1973, 1974).

Later Gold, Parizek, and Alexander (1973) proposed the concept of the "permancy of master streams" (major transverse drainages) when the number, magnitude and significance of lineament-related structures became apparent following more detailed study of LANDSAT and SKYLAB imagery and U-2 underflight photographs. Zones of fracture concentration revealed by lineaments and fracture traces were commonly associated with transverse drainages if not revealed by them, more productive well sites, foundation problems, etc. Vertical planes of structural weakness allow master streams to be let down across beds of variable resistance through geological time until these nearly vertical structural planes play out with depth or until streams are diverted through piracy. The fact that lineaments are straight in map view despite irregularities in topography, rock type or structure, demands that these planes are vertical to near vertical and may have depth influences approaching their length. Hence, master streams might be expected to remain fixed in map space as hundreds if not thousands of feet of rock are eroded.

This vertical plane master stream erosion model also works on the local tributary scale depending upon the size and length of the linear structural features and drainage area available to collect surface runoff. The number of lineaments in the 5 to 20 km or so length class as well as fracture trace class (< 1 km) far exceed the number used by transverse drainages of any size. See, for example, the lineaments map of Pennsylvania prepared by Kowalik and Gold (1976). Not enough water is available to establish major drainages along each lineament or minor drainages on fracture trace related structure. A stocastic process may have been involved in the evolution of transverse drainage networks where streams used some and not other zones of fracture concentration. The Tyrone-Mt. Union (Canich, 1976) and McAlevys Fort-Port Matilda lineaments (Hunter and Parizek, 1979) are both 45 km or longer in length, and yet only the Tyrone-Mt. Union lineament control major drainage, i.e., the Little Juniata River (Figure 5). The McAlevys Fort-Port Matilda lineament controls an underdrained karst valley of minor proportions by comparison (Hunter and Parizek, 1983) in Nittany Valley and a poorly developed watergap in Tussey Mountain. This watergap is confined to the inner ridge developed in the Oswego Sandstone. Only minor surface sags are developed in the Tuscarora Formation on both Tussey and Bald Eagle Mountains along this lineament. Its extension into Devonian to Pennsylvanian aged rocks is marked by a pronounced valley that extends into the Allegheny Plateau more or less parallel to Route 322. The lower reaches of Spring Creek and its well developed tributary, Logan Branch, on the other hand, and Nittany Creek all transect structure. Spring and Nittany Creeks cross Bald Eagle Mountain at Milesburg and Curtin Gaps. The surface expression of these lineament-related drainages are well developed (Figure 5) but both structures are short (15 miles) by comparison to the McAlevys Fort-Port Matilda lineament which is approximately 44 miles long.

The intensity of structures underlying lineaments vary with depth. Most likely some transverse drainages controlled by them abandoned their courses as the landsurface was lowered through time and more resistant rocks were encountered. Logan Branch may represent such a drainage that has lost its status. Its linear structural control is obvious between Gap Run along the western flank of Nittany Mountain where it cuts through the Oswego Sandstone. It extends across Nittany Valley, Bald Eagle Mountain and into the Allegheny escarpment to the west. Many other transverse drainages also extend up the Allegheny escarpment that define lineaments.

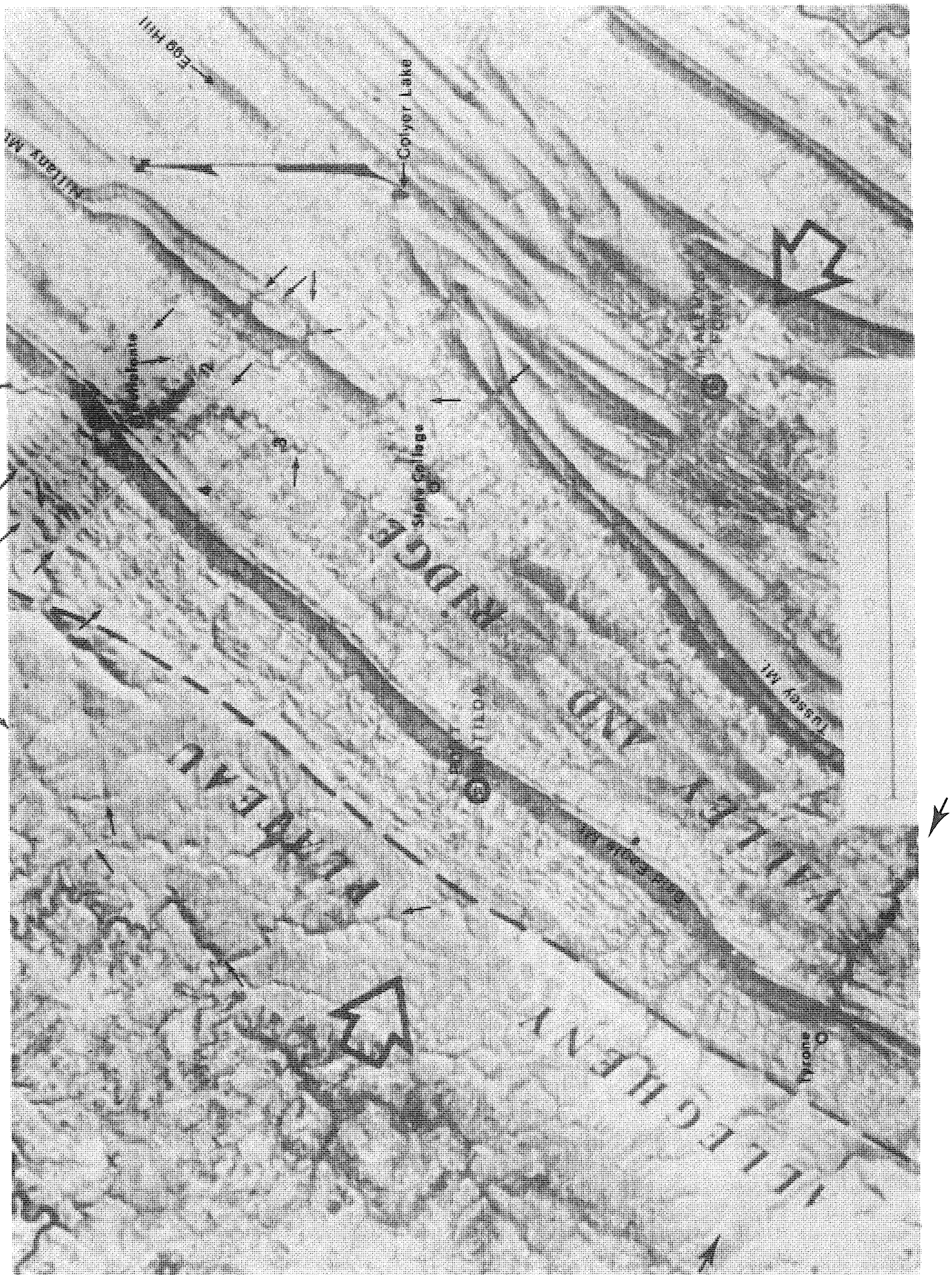


Figure 5. Selected lineaments in Nittany and adjacent valleys: 1, Bald Eagle Creek; 2, Logan Branch; 3, Spring Creek; 4, Buffalo Run; 5, Little Juniata River-Mt. Union-Tyrone lineament.

Logan Branch heads at 7 springs just below the village of Pleasant Gap and down slope from Gap Run that disappears in a swallow hole (Stop 8). Its zig-zag course across stratigraphic and structure strike within Nittany Valley is controlled by shorter fracture-trace related structures. Spring Creek near Lemont on the other hand, has only minor springs where it cuts across conduit prone limestones of Middle Ordovician-age. A water table map at the base of Nittany Mountain between Spring Creek and Logan Branch also reveals an interesting relationship. A water table trough extends 3.5 miles toward Spring Creek from the springs at the head of Logan Branch. The groundwater divide for this tributary groundwater basin is 3.5 miles from Logan Branch and only 1 mile from Spring Creek, the present master stream (Figure 6).

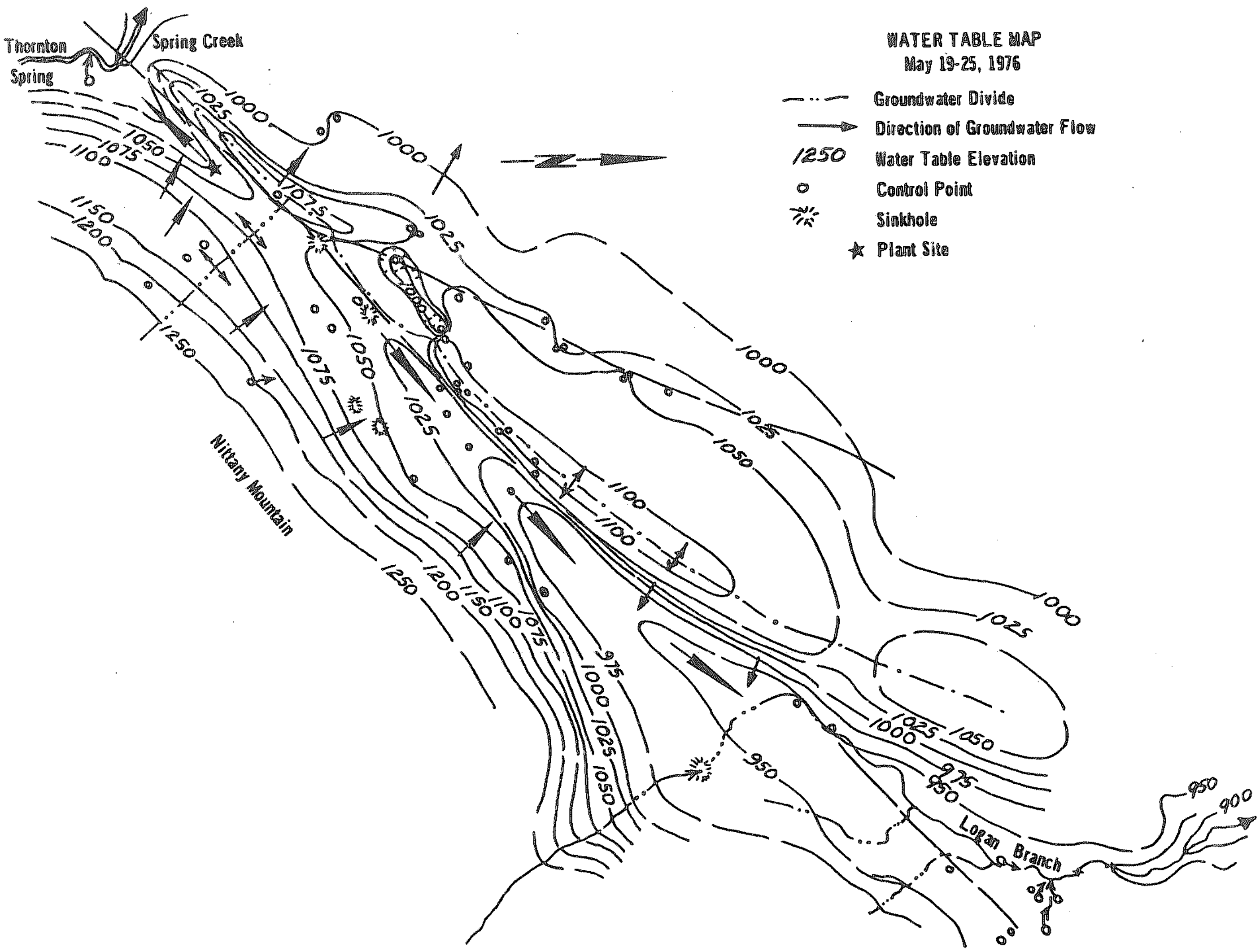


Figure 6. Water table configuration in the confluence between Logan Branch and Spring Creek.

These relationships suggest that Logan Branch was once the master stream and that its ancestral Nittany Mountain synclinal drainage and possibly Penns Valley drainage have long since been pirated by Spring Creek as it was able to erode through the plunging nose of Nittany Syncline. The groundwater basin along the western side of Nittany Mountain may still reflect the memory of Logan Branch's former more "glorious" status. Its groundwater basin to the northeast extends along

Route 64 beyond Zion for at least 7 miles. Details of the eastern segment of this groundwater flow system are obscured by an underground limestone mine that has resulted in a deep, localized cone of pumping depression. Marblehead Lime Company pumps up to 16,000 gpm to control groundwater flows during some periods of the year.

The relationship between stream valley location and limestone may be seen by reviewing various geologic and topographic maps of the area. Certain valleys appear to violate this generalization where they depart from limestone outcrop belts and cut across dolomitic strata. These cutoffs are almost always joint, zones of fracture concentration, or fault controlled. Valleys established across colluvium or alluvium at the base of ridges need not reflect these same structural and stratigraphical controls for short segments are consequent. These valleys segments have followed the consequent slope of Pleistocene colluvial-alluvial fans. Other valleys parallel the strike of limestone but are located in more resistant dolomite up dip (down stratigraphically) from the present limestone outcrop belt. This is accounted for by a space-temporal sequence of events illustrated in Figure 7.

When the drainage was first established, the land surface was higher as shown in Figure 7a. It was localized by differential solution along joints, bedding planes, more soluable beds and zones of fracture concentration within limestone. Integrated drainage developed directly above the limestone (Figure 7b) and down

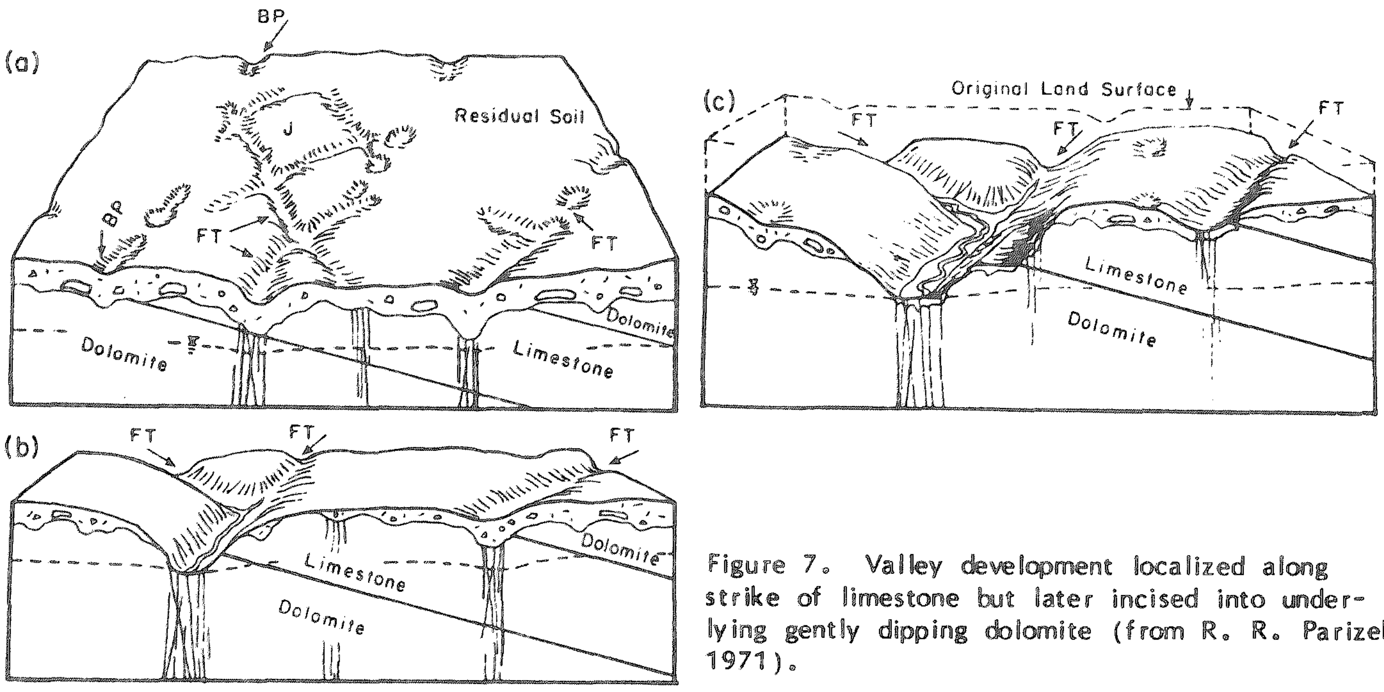


Figure 7. Valley development localized along strike of limestone but later incised into underlying gently dipping dolomite (from R. R. Parizek, 1971).

cutting and solution of land surface continued (Figure 7c) vertically along fracture zones, joints and faults which have fixed the channel position. The limestone outcrop belt continue to migrate in the down dip direction as the land surface was lowered. In time, the valleys may be incised completely into underlying dolomite and the initial limestone control valley location is no longer apparent. For larger drainages, valley widening is favored in the dip direction or toward the limestone valley wall.

It may be possible from these observations to estimate the amount of regional denudation which may have occurred since the valley position was fixed in plan view

(Figure 7c). This initial surface on which the drainage was established must have been graded to water gaps cut in resistant sandstone and quartzite which are responsible for maintaining local base levels for streams and groundwater. It can be shown that selected straight valley segments along Spring and other creeks are controlled by zones of fracture concentration revealed by lineament and fracture trace scale structures. The valleys show abrupt to nearly right angle turns at fracture trace intersections which produce an apparent meander pattern for this youthful, incised valley (Figure 8). From 50-90 percent of selected stream segments can be shown to be localized by these features in Nittany Valley. This is particularly evident along headwater tributaries where the control for valley location is still apparent. In adjacent uplands, solution along zones of fracture concentrations allows for subsidence of unconsolidated (residue) soils and produces lines of shallow depressions one to over ten feet deep (between HRB-Singer and the Kentucky Fried Chicken Restaurant on Route 322). Surface runoff concentrates in these depressions usually during the spring melt and after heavy rains. New tributary drainages are developed adjacent to valley walls as runoff becomes integrated. New straight valley segments are born in this manner where the water table stands near the land surface. To a lesser degree, master joint sets control the same process and together with zones of intersecting fracture concentrations and bedding plane openings, they produce irregular offsets in the new tributaries (Figure 8). Normal faults produce similar controls on drainage near State College (Parizek, 1971).

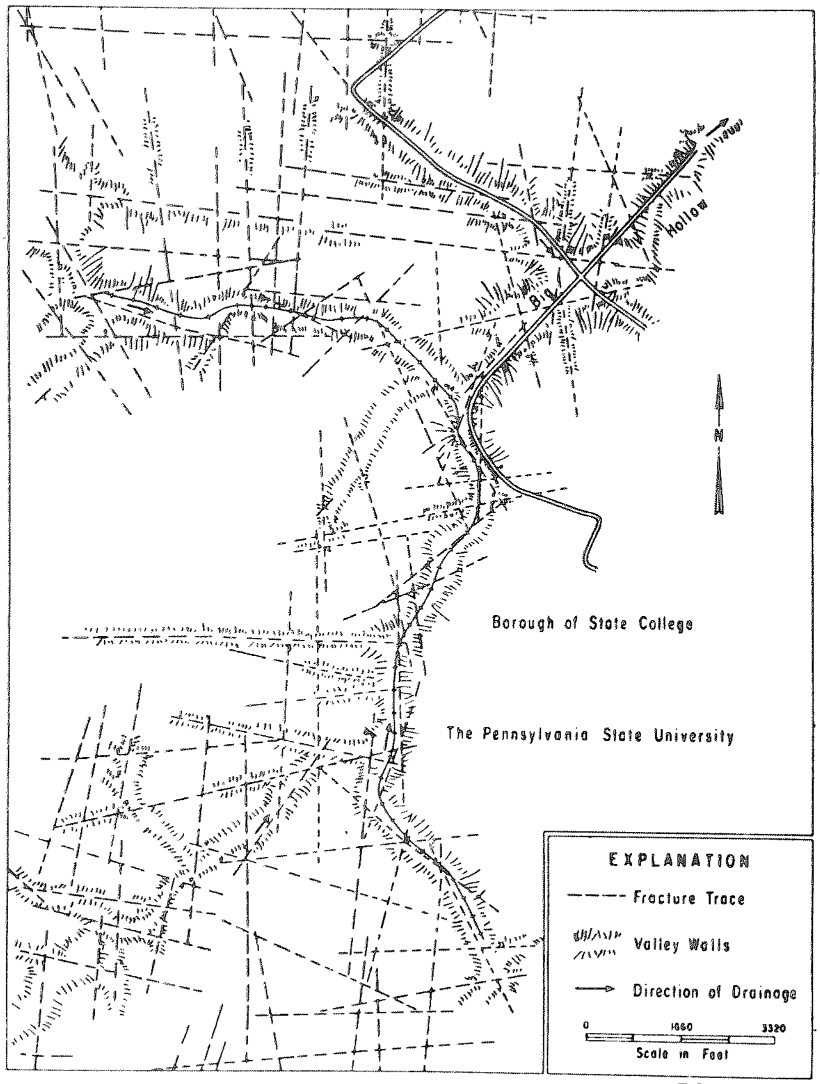


Figure 8. Valley development localized by zones of fracture concentration (from R. R. Parizek, 1971).

Subsurface Drainage

Many of the tributary valleys are underdrained and lack throughgoing surface streams. Sink holes are common along these valleys. Dolomite normally underlies uplands that have residual soil covers of varying thickness, 1 to 300 ft. Although these sediments may be depressed into solution zones in the top of bedrock, open sink holes are not common. Near surface cavities in limestones by contrast normally are larger and include both processes of solution and collapse. Internal drainage networks soon are established in limestone because of its increased solubility. Residual soil deposits are readily transported into sink and swallow holes and carried away along subsurface channel ways. Outcrops in limestone are more numerous when compared to outcrops observed for adjacent regions underlain by some dolomites. A topography may result where the limestone protrudes above adjacent, less soluble dolomite. This may be accounted for as follows. Once subsurface drainge is established within limestone, solution is concentrated in sink and swallow holes, and subsurface channel ways. Where the downcutting process is fast enough as at the heads of tributaries, undissolved masses of limestone remain only to be dissolved by direct precipitation and subsurface lateral solution. By contrast, solution openings along bedding planes and joints within dolomite are smaller and not as well interconnected. Their residual soils cover wide areas of bedrock. Surface water infiltration varies in topographically high and low areas according to detailed variations in soil characteristics and drainage. Surface runoff redistributes a greater proportion of precipitation to surface depressions where soil may be two to three times thicker than above adjacent topographically high areas. Percolating soil water rich in CO_2 tends to move through the entire soil profile attacking the top of bedrock more uniformly. The net effect is that the entire dolomitic landscape is lowered faster than some adjacent limestone areas now largely devoid of soil. Various examples of this erosion rate difference can be seen in central Pennsylvania. On the other hand, where slightly more resistant shale or sandstone is present (Salona-Coburn Formations, Upper Sandy Dolomite Member, Dale Summit Sandstone) or more dense dolomite (Tea Creek Member), these areas are marked by uplands of low relief (Parizek, 1971).

This seemingly anomalous weathering may be accounted for when one considers the role of the soil cover. Carbonation occurs in the soil in excess of what is possible in the atmosphere by at least an order of magnitude. Soil waters enriched in CO_2 are capable of dissolving more carbonate minerals than an equal volume of water which has fallen on exposed limestone. Once the soil has been washed into subsurface karst openings, solution lowering of bare rock surface will be slowed. Solution will continue along rapidly developing subsurface passageways developed around and below limestone pinnacles (Parizek, 1971).

Once the tributary drainages are established above limestone, the "cannibalistic process" of localized solution beneath the topographic low is speeded up. Surface water undersaturated with respect to $CaCO_3$ concentrates along the growing valleys at the expense of adjacent uplands. Permanent streams may be established for a time once valleys are deepened to the water table until subsurface solution zones may develop and underdrain these valleys. This further favors increased solution within the valley environment. The end result is to produce (1) an aquifer with major, regionally interconnected solution openings; (2) cavities and channel ways containing increased amounts of clay, sand, and gravel reworked from the initial unconsolidated overburden; (3) valley walls, floors, and adjacent uplands with discontinuous patches of residue soil cover, hence regions susceptible to rapid infiltration and to pollution; (4) regional groundwater underdrains producing 10-100 ft. of local relief on the free water surface; (5) sinking creeks, hence major groundwater recharge areas capable of accepting million of gallons of water a day; (6) permanent surface drainages; and (7) conduits and more permeable aquifers where groundwater flow rates are rapid (Parizek, 1971).

Fragments of the conduit drainage system can be observed by examining the caves that occur widely in Nittany Valley. Some 90 caves are known in Centre County (Dayton and White, 1979) and additional caves are located in the same drainage basin in adjacent Clinton, Huntingdon, and Blair Counties. Caves, particularly Pennsylvania caves, are small and usually represent only tiny fragments of the present- or paleo-drainage system. The amount of cave passage explored in Nittany Valley is far too short to establish much of the paleodrainage or to establish the boundaries of the present-day groundwater basins.

Three sorts of caves are found. There are caves associated with the swallow holes where the mountain streams sink at the limestone contact. Noll Cave (Figure 9) is part of the floodwater inlet, now choked with sediment, at the sink of Gap Run (Stop 8). Stream sink caves are often used by the contemporary streams in their lower levels while higher levels represent abandoned channels or flood overflow routes. Sharer Cave and Veiled Lady Caves in Penns Valley are other examples. These caves often have high gradients consistent with the steep gradients of the mountain streams that feed them. The second sort of caves are found at spring mouths and allow the drainage system to be explored upstream into the subsurface. Few springs in Nittany Valley are explorable but the commercial Penns Cave is an

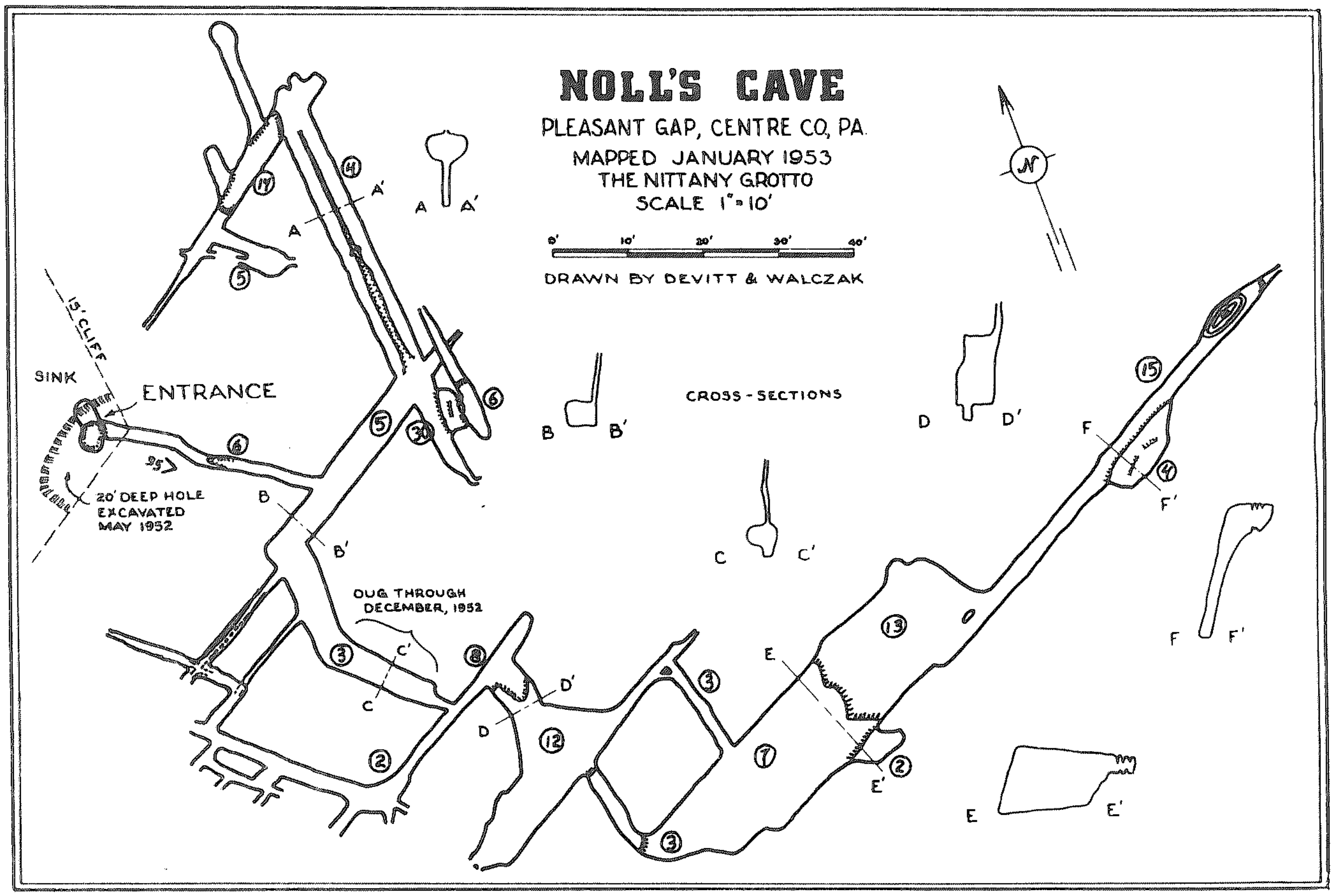


Figure 9. Noll Cave lies beneath a choke of sediments several hundred feet upstream for the swallow hole of Gap Run. High sediment loads carried by long-return period floods have a tendency to choke cave entrances in stream sink environments.

exception. The spring mouth at Penns Cave was not originally an explorable opening; the natural entrances are through sinkholes upstream along the master trunk channel. The third sort of caves are the fragments of old conduit now abandoned because of retreat of underground drainage to lower levels. These come in many sizes and shapes. Some are clearly related to present streams and can be thought of as a sort of underground terrance level. Others are related to the valley upland surface itself.

Figure 10 shows the distribution of caves and cave passage lengths for 75 caves in the Nittany Valley drainage. In a superficial way it appears that there is the greatest concentration of caves at the most probable elevation but there is more to it. The concentration of cave passage in the 800-850 foot interval occurs near the Juniata River and represents a major "terrace" 100-150 feet above the River. The cluster of caves and cave passage development between 900 and 950 feet are mostly in the lower reaches of the drainage system and are spaced about 100 feet above the present stream profile. The large cluster of caves between 1000 and 1100 feet is more problematic but could be related to the level developed near Houserville on the Spring Creek Drainage (Stop 5). Finally there is a large concentration of caves and cave passages near 1200 feet with fewer caves and less passage tapering off at higher elevations. These include some of the largest caves in the valley and also probably the oldest since they relate to the valley upland surface.

Stream alluvium derived from nearby mountain ranges becomes trapped in these growing sink holes and is deposited in caves, conduits, and other openings distant from swallow holes. Other sediment infiltrates from the valley floor through sinkholes and solutionally widened fractures. Conduit fillings range in texture from clay, silt, sand and gravel with pebbles up to 2 1/2 inches or more in diameter 2000 ft. or more from the nearest cave or solution opening. Assorted organic matter and clastic sediments may be present. Shale, quartzite, sandstone and chert pebbles to 2 1/2 inches in diameter have been recovered from gravel-filled conduits in the Bellefonte Dolomite and Middle Orodivician limestone at depths to 150 ft. below the water table. This is in an area where the water table has never been lower than at present. These caves and conduits formed at least 100 ft. below the water table. Runoff from mountain watersheds and watergaps may approach thousands of gallons a minute during periods of snowmelt. This water plunges into sink holes developed downslope from the Reedsville Shale and Salona-Coburn Formation and may cascade 30 to 150 ft. to the underlying water table. It is easy to see how groundwater flow velocities quickly mount to well above turbulent levels allowing transport of cobbles and boulders and how undersaturated groundwater becomes available well below the water table and distant from points of recharge. These factors favor concentrated and rapid solution of mountain base carbonates at the expense of adjacent carbonates. Some conduits transmit water only during wet-weather and extreme flood events. Others are well below the permanent water table.

The tendency for caves to maintain a horizontal longitudinal profile in spite of structure and lithology is accounted for mainly by the position of the water table. Once an integrated system of openings becomes established, it is capable of draining vast quantities of groundwater. Very low gradients are sufficient to convey all available groundwater to discharge areas even during peak periods of runoff. As the drainage net and groundwater basin becomes larger, conduits become better integrated and fewer in number.

Once these networks become established, chemically aggressive waters follow these avenues and dissolve strata within the region of water table fluctuation.

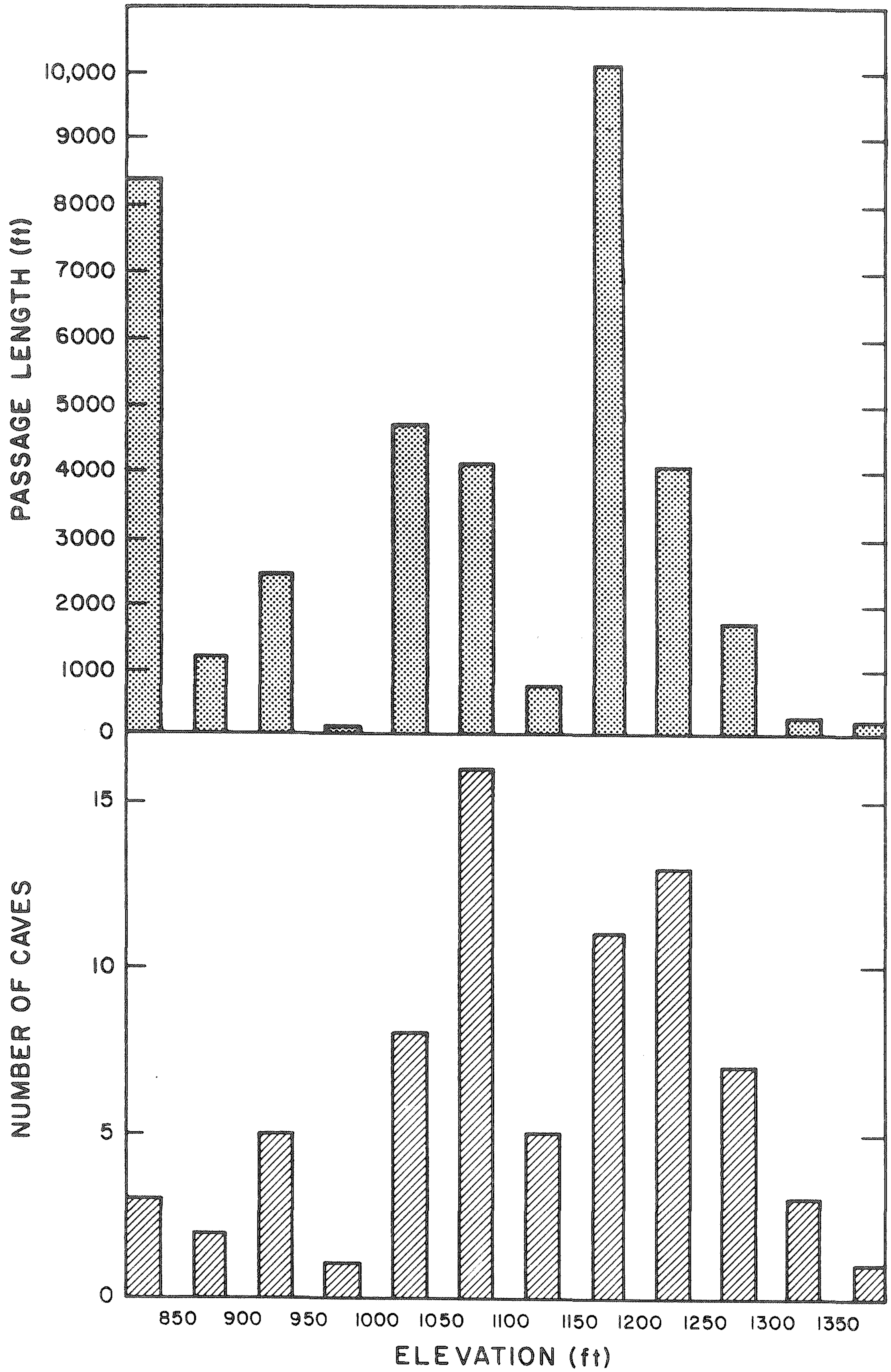
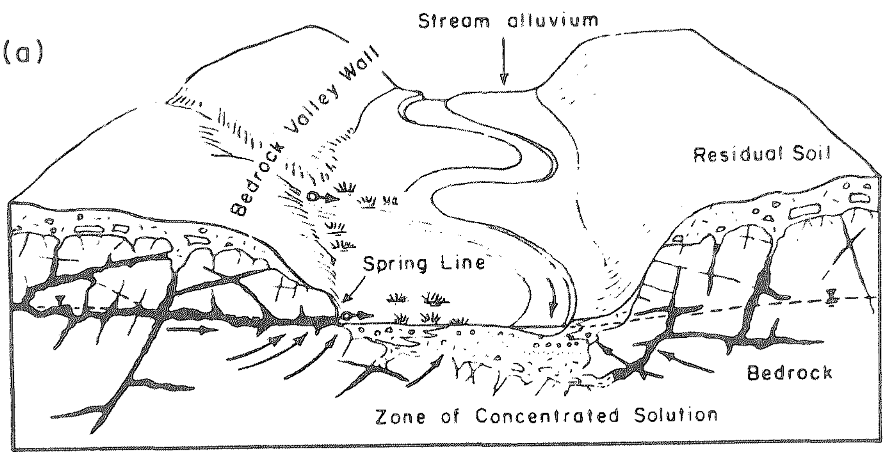


Figure 10. The distribution of passage length and number of caves in 50 foot elevation intervals through Nittany Valley. Data drawn from the Pennsylvania Cave Data Base maintained at the Pennsylvania State University by K. Wheeland.

Solution at greater depths below the water table is possible beneath regional recharge areas and where concentrated recharge occurs in swallow holes. Large solution openings may extend to greater depths below the water table (150 - 300 ft.) in these areas. Data from caliper logs indicate that cavities 1 to 5 ft. high can occur locally at depths of 500 ft. below ground level and 300 to 400 ft. below the water table.

The influence of water-table position on the solution process can be seen best at groundwater discharge areas. Springs most frequently emerge between bedrock valley walls and the contact with colluvial-alluvial valley fill sediments (Figure 11). These zones of groundwater discharge are usually at or slightly below the water table, are graded to the valley floor, and reflect zones of increased solution which extend beneath adjacent uplands at a gentle slope. Where valleys subsequently have been underdrained, these shallower conduits become abandoned. Here, younger and deeper conduit systems develop which are graded to underlying groundwater drains located beneath the valley floor. Wet weather springs occasionally mark the outlets



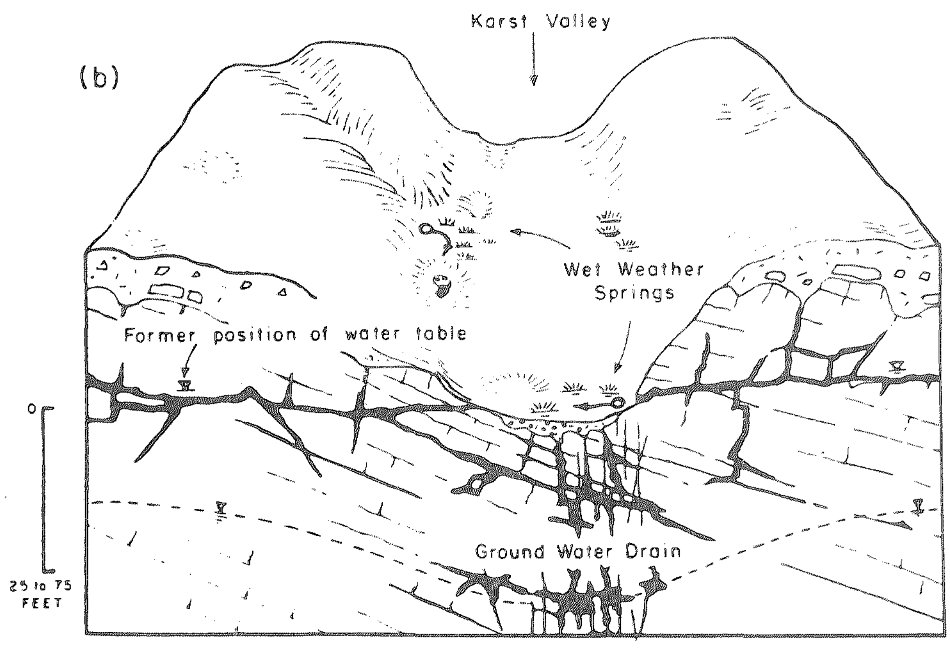


Figure 11. Early stages in karst valley development (from Parizek, 1971).

of these otherwise abandoned conduit systems. Where valleys are just now becoming underdrained (Stop 6) but still contain appreciable intermittent surface runoff, permanent and wet weather springs may still be present usually at the contact between the valley wall and less permeable valley fill sediments (Figure 11).

If cave levels, as indicated in a rough way by Figure 10, represent old terrace levels, some record of the chronology of the valley should be recorded in the caves. Although the cave sediments are complex sequences of sands, silts, and gravels, nothing equivalent to a stratigraphic section has been constructed. The depositional processes are chaotic and vary greatly from one part of a cave to the next. In spite of a rich sedimentary record, little use has so far been made of it.

Flowstone and dripstone deposits can be dated by $^{234}U/^{238}U$ and $^{230}Th/^{234}U$ disequilibrium with a time scale that extends to about 350,000 years BP (Thompson et al., 1975; Harmon et al., 1978). Most travertines are considerably younger than the caves in which they are deposited but dating of travertines from West Virginia caves show two clusters of dates: ages less than 6000 years BP and ages in the range of 70,000 to 200,000 years. These should represent post-Wisconsinian carbonate deposition and carbonate deposition during the Sangamon interglacial. It appears from these and other studies that carbonate deposition almost shuts down during the glacial advances. There is no uranium age data for Pennsylvania caves but visual inspections of dripstone deposits in some of the more highly decorated caves such as the commercial Lincoln Caverns reveals what appears to be two populations of travertine. There are underlying massive deposits overgrown with much smaller and somewhat differently colored deposits. It is suspected that the massive deposits are the Sangamon age material, largely because of a longer period available for growth and the small overgrown speleothems represent post-Wisconsinian deposition.

The clastic sediments in caves do contain minor amounts of magnetic minerals which provide a paleomagnetic signal. The iron oxides present are not strongly magnetic, they are few in number, and they are aligned only in a statistical sense because of turbulence and buffeting by sand and silt during deposition, but the magnetic signal can be measured. In a pioneering study of sediments in Mammoth Cave, Schmidt (1982) was able to follow the pattern of magnetic normals and magnetic reversals to 2 million years before present in the highest levels of the cave, thus confirming geomorphological evidence that the oldest part of the cave system are late Pliocene or very early Pleistocene age. Similar measurements have not been made in Pennsylvania caves but there is nothing inherently impossible in the hypothesis that the highest caves levels, those associated with the valley upland erosion surface, are of pre-Pleistocene age.

If the base level is fixed for a prolonged period, denudation dominated by dissolution within the vadose zone progresses and the land surface approaches the water table. Where this condition obtains, secondary porosity and permeability development tends to be limited with depth. Favorable aquifer characteristics are normally restricted to depths of 50-300 ft. below land surface and may be generally less than 150 ft. Where residual soils are available, the denudation process may continue somewhat uniformly almost to the water table.

Regions with extensive, shallow water tables and poorly drained lowlands reflect this condition. Examples may be seen near Chambersburg, Pennsylvania and Fredrick and Hagerstown, Maryland, and Martinsburg, West Virginia, and similar other regions where the landsurface has been eroded nearly to the local base level and the water table stands within a few feet to only several tens of feet below landsurface.

Here, deep internal drainage is restricted. Residual soils that must have existed previously similar to what are observed in Nittany and adjacent western Appalachian valleys have been largely eroded away to expose numerous bedrock outcrops. Aside from the shallow position of water table, reduced internal drainage hence, increased surface runoff and erosion, the general lack of a thick regolith in eastern carbonate valleys in the Ridge and Valley may also be due to their close proximity to master streams and regional base level. More time was available to permit erosion of residual soils approaching the eastern seaboard because these rivers would have been cut to base level for longer periods of time, and the water table would have been adjusted to baselevel earlier than for western interior basins.

Soils

The soils of the unglaciated Ridge and Valley show a strong association to landform (Figure 12, Table 1). Marchand et al. (1978) estimated that on average, soils formed on residum occupy approximately 67% of the Ridge and Valley. These occur on the more gently rolling valley floors and uplands while soils formed on colluvium occupy approximately 27% of the area. Colluvium occur on the lower one-half to three-fourths of the side slopes of major and secondary ridges (Stops 7 and 8). Colluvium also is widespread within the valley upland setting in Nittany Valley along intermittent drainage ways, and on the sides of smaller hills. The remaining 6% of the area is occupied by flvuial deposits including floodplain and terrace deposits (Ciolkosz et al., 1979). Aside from showing a strong association to landform, these soils are strongly related to parent material. For this reason, Ciolkosz et al. (1979) arranged Pennsylvania colluvial soils into four parent material-drainage groups and residual soils into seven classes (Table 1). They regard this sequential arrangement as a natural association of these soils in the landscape.

Residual Soil

The great bulk of unconsolidated overburden deposits that blankets carbonate bedrock in Nittany Valley is classified as residual having formed from prolonged physical and chemical weathering. These deposits have been documented to be at least 365 feet thick locally (just southwest of Gameland 176) and to vary with rock type and topographic setting. Bedrock outcrops are most abundant for the Bellefonte Dolomite, especially the Tea Creek Member which is a dense, crypto-crystalline dolomite. This unit is rather pure hence does not produce thick residual soil, nor does it favor internal or differential weathering of the top of bedrock. Its ridge forming character together with its low insoluable residue content and low permeability favors surface runoff and soil erosion.

The thickest residual soil deposits occur above the Gatesburg Formation especially the Mines Dolomite, and Upper and Lower Sandy Dolomite Members. These units yield higher proportions of insoluable residue when weathered, support a deep water table, are higher in bulk permeability and porosity than all other carbonate rocks in the region hence, promote internal drainage. Differential weathering is enhanced and detention storage is maximized by a myrid of well developed closed depressions.

Broad areas of the valley upland contain thick residual soils (sapolites or regolith) especially at interfluves. There is much evidence for their residual character (Stops 3 and 4).

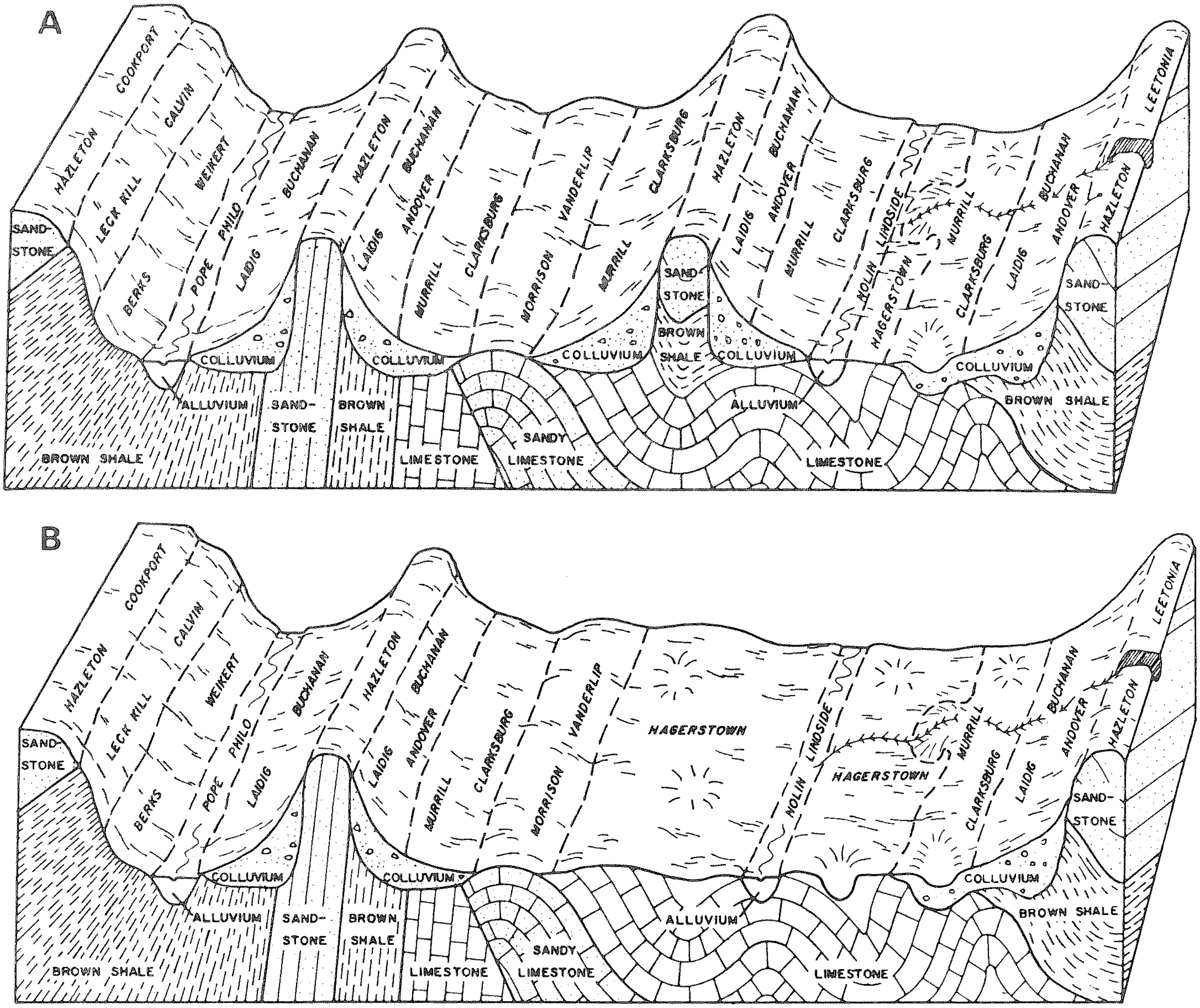


Figure 12. Soil-landscape relations of Nittany Valley (A is northeast and B is southwest of State College) (from Ciolkosz et al., 1980).

(1) Bedding and lithologic variations in the soil can be matched with individual bedrock units from which they were derived.

(2) Structures in soil are relict and can be matched with similar structures in bedrock, i.e., mud cracks, ripple marks, spacing and relative abundance of insoluable residue, etc.

(3) Residual beds of chert, quartzitic sandstone, oolitic chert and sandstone can be matched with bedrock parent material.

(4) Rock fragments tend to be angular, nodular or irregular in shape not rounded, sorted or otherwise indicating that they have been transported by wind, water, gravity or ice.

(5) Soil units often can be correlated between drill holes or traced within trench excavations. These beds do not display bedforms or overall character of having been transported and deposited within fluvial, lacustrine or related environments.

Table 1 Major Soils of Central Pennsylvania Arranged According to Parent Material and Drainage.*

Parent Material	(Shallow) <20" to bedrock	(Moderately Deep) 20-40" to bedrock	(---- Deep >40" to consolidated bedrock ----) Drainage Class and Depth to Mottling (---- Well Drained ----) (---- >40" ----)	Moderately Well Drained (20-40")	Somewhat Poorly Drained (10-20")	Poorly Drained 0-10"; some gleying)	Very Poorly Drained (0-10"; strong gleying)
Residual							
Gray and brown acid shale and siltstone	Weikert Lithic Dystrochrept; loamy-skeletal	Berks Typic Dystrochrept; loamy-skeletal	Bedington Typic Hapludult; fine-loamy Hartleton Typic Hapludult; loamy-skeletal	Blairton [Mod. deep------] Aquic Hapludult; fine-loamy Comly--------------------) Typic Fragiudalf; fine-loamy		Markes [Mod. deep] Typic Ochraqualf; loamy-skeletal Brinkerton Typic Fragiaqualf; fine-silty	
Gray and brown acid shale siltstone and some clay shale	Weikert Lithic Dystrochrept; loamy-skeletal	Gilpin Typic Hapludult fine-loamy	Rayne Typic Hapludult; fine-loamy	Wharton Aquic Hapludult; clayey	Cavode Aeric Ochraquult; clayey	Armagh Typic Ochraquult; clayey	
Red acid shale and siltstone; dull red 4 chroma or less	Klinesville Lithic Dystrochrept; loamy-skeletal	Calvin Typic Dystrochrept; loamy-skeletal	Leck Kill Typic Hapludult; fine-loamy	Albrights----------------) Aquic Fragiudalf; fine-loamy		Conyngham Unclassified	
Gray and brown acid sandstone	Ramsey Lithic Dystrochrept; loamy-skeletal †	Dekalb Typic Dystrochrept; loamy-skeletal	Hazleton Typic Dystrochrept; loamy-skeletal Clymer Typic Hapludult; fine-loamy	Cookport-----------------) Aquic Fragiudult; fine-loamy		Nolo Typic Fragiaquult; fine-loamy	Lickdale Humic Haplaquept; fine-loamy
Red acid sandstone; dull red 4 chroma or less	Ramsey Lithic Dystrochrept; loamy-skeletal	Lehew Typic Dystrochrept; loamy-skeletal	Ungers Typic Hapludult; fine-loamy	Albrights----------------) Aquic Fragiudalf; fine-loamy		Conyngham Unclassified	
Grayish brown sandstone (in some places a very sandy limestone)			Vanderlip Typic Quartzipsamment; coated Morrison Ultic Hapludalf; fine-loamy Leetonia Entic Haplorthod; sandy-skeletal, siliceous Gatesburg Entic Haplorthod; sandy, siliceous				
Very cherty limestone			Elliber Typic Hapludult; loamy-skeletal	Kreamer Aquic Hapludult; clayey	Evendale-------------------) Aeric Ochraquult; clayey		
Cherty limestone			Hublersburg Typic Hapludult; clayey				
Relatively pure limestone	Opequon Lithic Hapludalf; clayey		Hagerstown Typic Hapludalf; clayey**	Clarksburg Typic Fragiudalf; fine-loamy	Penlaw Aquic Fragiudalf; fine-silty	Thorndale Typic Fragiaqualf; fine-silty	

Parent Material	(Shallow) <20" to bedrock	(Moderately Deep) 20-40" to bedrock	Deep >40" to consolidated bedrock — Well Drained >40"	Moderately Well Drained (20-40")	Somewhat Poorly Drained (10-20")	Poorly Drained (0-10"; some gleying)	Very Poorly Drained (0-10"; strong gleying)
Residual (cont'd.)							
Thin bedded limestone and calcareous shale		Ryder Ultic Hapludalf; fine-loamy	Duffield Ultic Hapludalf; fine-loamy Edom Typic Hapludalf; clayey**, illitic Frankstown Typic Hapludult; fine-loamy	Clarksburg Typic Fragiudalf; fine-loamy	Penlaw Aquic Fragiudalf; fine-silty	Thorndale Typic Fragiaqualf; fine-silty	
Colluvium							
Brown and gray acid shale, siltstone and fine grain sandstone			Shelocta Typic Hapludult; fine-loamy	Ernest Aquic Fragiudult; fine-loamy	------)	Brinkerton Typic Fragiaqualf; fine-silty	
Red acid shale, siltstone and fine grain sandstone; dull red chroma 4 or less			Meckesville Typic Fragiudult; fine-loamy	Albrights Aquic Fragiudalf; fine-loamy	------)	Conyngham Unclassified	
Gray and brown acid sandstone and shale			Laidig Typic Fragiudult; fine-loamy	Buchanan Aquic Fragiudult; fine-loamy	------)	Andover Typic Fragiaquult; fine-loamy	
Brown and gray limestone, shale and sandstone			Murrill Typic Hapludult; fine-loamy	Clarksburg Typic Fragiudalf; fine-loamy	Penlaw Aquic Fragiudalf; fine-silty	Thorndale Typic Fragiaqualf; fine-silty	
Very cherty limestone and shale			Mertz Typic Hapludult; loamy-skeletal	Kreamer Aquic Hapludult; clayey	Evendale Aeric Ochraquult; clayey	------)	
Recent Alluvium (Floodplains)							
Alluvium from acid gray and brown shale, siltstone and sandstone uplands.			Pope Fluventic Dystrochrept; coarse-loamy	Philo Fluvaquentic Dystrochrept; coarse-loamy	Stendal Aeric Fluvaquent; fine-silty	Atkins Typic Fluvaquent; fine-loamy	Elkins Humaqueptic Fluvaquent; fine-silty
Alluvium from acid red shale, siltstone and sandstone uplands; dull red chroma 4 or less			Barbour; Linden Fluventic Dystrochrept; coarse-loamy	Basher Fluvaquentic Dystrochrept; coarse-loamy	------)	Holly Typic Fluvaquent; fine-loamy	Papakating Mollic Fluvaquent; fine-silty
Alluvium from limestone, shale and siltstone upland			Nolin Dystric Fluventic Eutrochrept; fine-silty	Lindside Fluvaquentic Eutrochrept; fine-silty	Newark Aeric Fluvaquent; fine-silty	Melvin Typic Fluvaquent; fine-silty	Dunning Fluvaquentic Haplaquoll; clayey**
Old Alluvium (Terraces) and Lacustrine							
Gray and brown acid shale siltstone and sandstone			Allegheny Typic Hapludult; fine-loamy	Monongahela Typic Fragiudult; fine-loamy	Tyler-Aeric Fragiaquult; fine-silty Tygart - Aeric Ochraquult; clayey	Purdy Typic Ochraquult; clayey	

* Almost all soils are also mixed, mesic.
** These soils are classified in the fine family but for the purpose of this table clayey will be used.
In Pennsylvania most pedons are skeletal.

(6) The mineralogy, texture and character of float, etc. within the overburden can be traced along stratigraphic strike and correlated with underlying bedrock units from which it was derived. Foreign rock fragments are absent.

(7) The contact between the regolith and underlying bedrock can be gradational characteristic of a sapolite.

(8) Distinctive soil units occupy distinctive positions on the landscape, i.e., sandy soils overlie the Upper Sandy Dolomite Member of the Gatesburg Formation, thick kaolinite deposits occur above and along the Lower Sandy Dolomite Member, irregular masses of oolitic chert follows the Mines Dolomite Member of the Gatesburg, some soil units show distinctive color above the Trenton series of carbonates, etc.

Figures 13 and 14 show residual soil thickness at two sites near State College. Both are underlain by the somewhat sandy Gatesburg Formation. The maximum thickness of residual soil observed was in coreboring GM-6. Split spoon samples were collected at 2.5 foot intervals to a depth of 162.5 feet before the first bedrock was encountered. Much of the Gameland 176 contains in excess of 100 feet of residual soil. Residual soils at the Agronomy-Forestry Waste Water Treatment area exceed 60 feet in thickness but thin along the upland edges of tributary valleys where more extensive surface erosion has occurred. Bedrock outcrops are restricted to valley walls and valley bottoms where residual soils have been totally removed.

Even greater thicknesses of soil were noted in test wells drilled above the Lower Sandy Dolomite Member of the Gatesburg Formation. Two air rotary holes drilled on the Stevenson Property by the Borough of State College (Water) Authority penetrated regolith to depths in excess of 365 feet before any resistant bedrock was encountered. This is the thickest residual soils documented in Nittany Valley. It occurs above inclined beds of the Lower Sandy Dolomite Member where bedrock dips may exceed 15 to 20°. Beds are not vertical to near vertical as observed at Stop 2.

Variations in soil thickness within a given area are related to rock characteristics, presence of systematic and non-systematic joints, their intersections, zones of fracture concentration, faults, and bedding plane partings and the amount of post dissolution erosion. Usually, soil thickness is a maximum below surface sags and depressions where internal drainage dominates over surface runoff. For example, soils may be 60 to 100 feet thick below closed surface depressions and only 5 to 20 or so feet thick above immediately adjacent topographic highs. The surface topography of internally drained uplands strongly mimics the bedrock surface topography underlying residual soil deposits.

Subsidence depressions can persist for very long periods of time, both at the landsurface and bedrock surface. Topographic evidence for the former can be eliminated if the rate of surface erosion exceeds the rate carbonate rock dissolution at the soil-bedrock contract. In time, topographic irregularities related to differential weathering of bedrock are truncated by surface erosion and eliminated. Where topographic evidence for former closed surface depressions have been lost, sites of maximum dissolution can still be observed by (1) mapping the configuration of the top of bedrock, (2) mapping letdown structures within the soil overburden revealed by lithologic variations inhereted from the bedrock, i.e., silt, sand and claybeds, chert horizons, etc., and (3) mapping sinkhole fills, etc.

Where lithologic variations are distinct within soil overburden deposits, beds mapped within topographic lows within the bedrock surface will rise in elevation approaching bedrock highs or may be exposed and truncated at the land surface. In extreme cases of soil collapse, evidence for relict bedding may be eliminated during this inversion process.

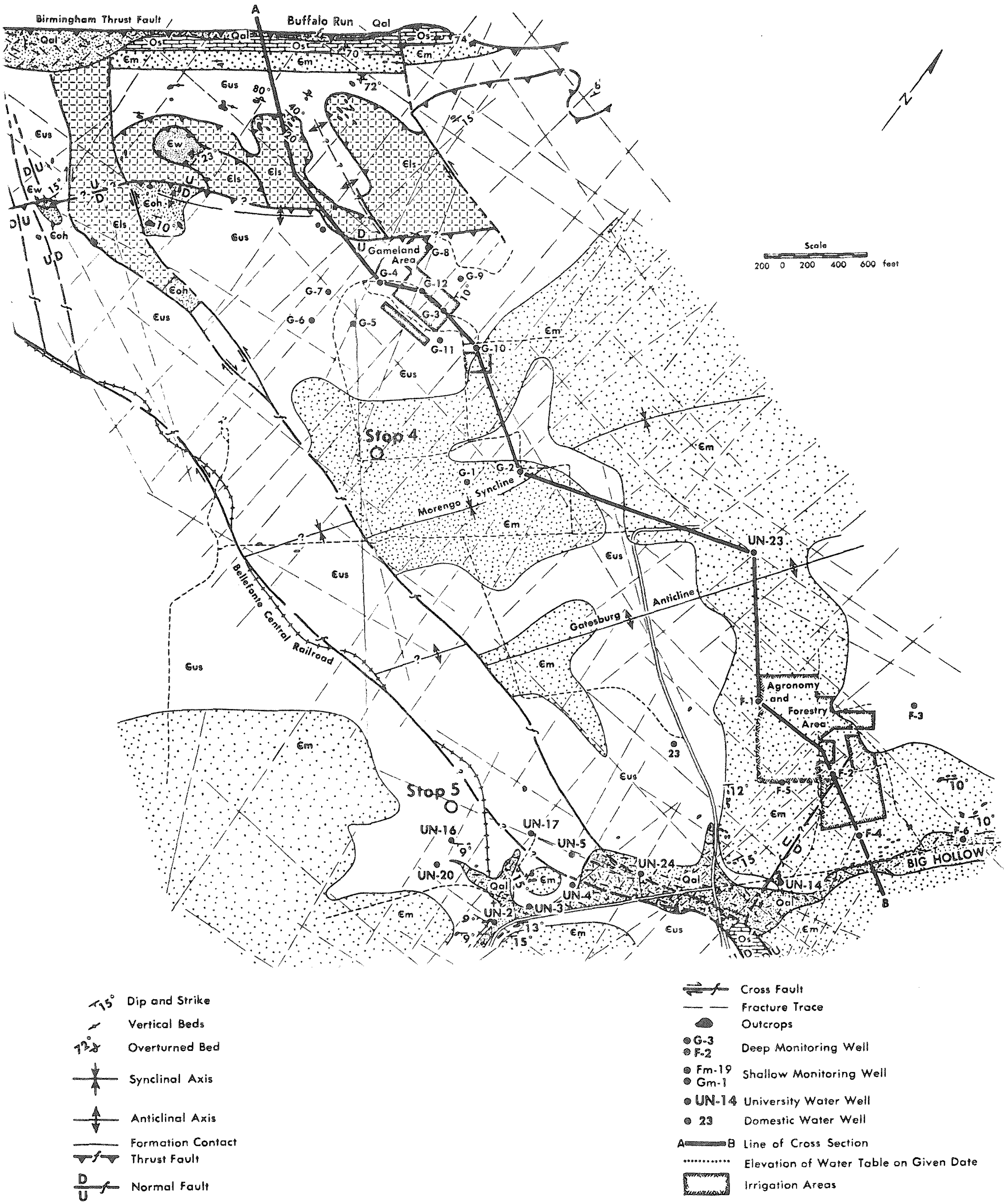
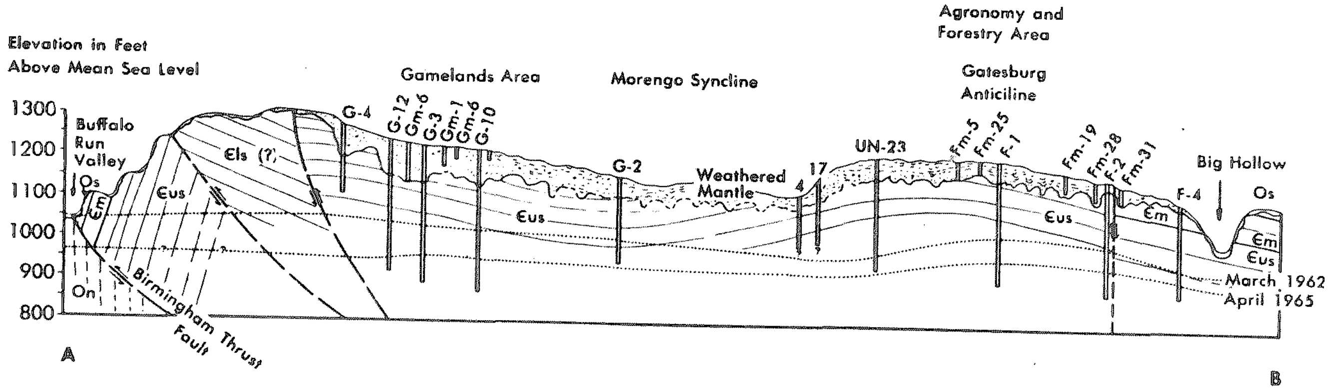


Figure 13. Geologic map of Penn State's Living Filter Project area (from Parizek et al., 1967).



STRATIGRAPHIC AND SOIL SUMMARY

Age		Description	Soil Series	Texture
Recent	Qal Quaternary Alluvium	Mixture of clay, sand, chert, and quartzite cobbles and boulders 1 to 30 feet	Araby, Huntington	Coarse to Medium
Lower Ordo.	On Nittany Dolomite	Thick bedded, coarsely crystalline dolomite, fine to coarse, chert nodules. 1209 ± feet.	Hagerstown-Hublersburg	Medium to Fine
Lower Ordo.	Os Stonehenge Limestone	Fine-grained limestone, some dolomite beds, edgewise conglomerate. 260 to 702 feet.	Hagerstown-Hublersburg	Medium to Fine
Upper Cambrian, Gatesburg Formation	Єm Mines Dolomite Member	Massive dolomite, beds of oolitic chert, sand stringers. 130 to 230 feet.	Hublersburg-Hagerstown	Medium to Fine
Upper Cambrian, Gatesburg Formation	Єus Upper Sandy Member	Recurring beds of dolomite, sandstone, and quartzite. 650 to 700 ± feet.	Morrison, Gatesburg	Coarse
Upper Cambrian, Gatesburg Formation	Єoh Ore Hill Member	Massive dolomite, beds of platy dolomite. 260 feet.	Hublersburg-Hagerstown	Medium to Fine
Upper Cambrian, Gatesburg Formation	Єls Lower Sandy Member	Recurring beds of dolomite, sandstone, and quartzite less abundant. 500 feet.	Morrison, Gatesburg	Coarse
Upper Cambrian	Єw Warrior Limestone	Limestone interbedded with some dolomite, thin shale units. 600 feet.	Hagerstown-Hublersburg	Medium to Fine

Figure 14. Geologic cross-section through Penn State's Living Filter Project area (from Parizek et al., 1967).

The significance of zones of fracture concentration revealed by fracture traces and lineaments in differential weathering of carbonate bedrock is dramatically shown in figure 15. Foundation borings were made for the East Halls dormitory complex on the Penn State University Campus. Borings and later buildings were located above two fracture trace-related structures and their intersection within the Stonehenge Limestone and overlying Nittany Dolomite. Physical and chemical weathering of the Nittany Dolomite were enhanced along these zones of fracture concentration. Depths to bedrock were 25, 50 to 75 feet on the structures and < 5 to 25 feet or so at immediately adjacent sites located off these structures. The cumulative thickness of weathering products is obvious and must include the insoluble residue derived from a significant volume of carbonate bedrock. The insoluble residue content of the Nittany Dolomite is lower than that measured for the Mines and Upper Sandy Dolomite Members of the Gatesburg Formation. Hence, its residual soils represent the remnants of weathering products resulting from the removal of several thousands of feet of carbonate bedrock that have accumulated slowly over millions of years.

High angle faults observed in the Gameland 176 area also localize thick residual soils that exceed 100 feet, surface sags and depressions and perched ponds all indicators of differential weathering of the underlying bedrock surface.

Kaolinitic Soils from Barrens

Feldspathic shales, siltstones and sandstones within the Gatesburg Formation have been identified by R. Pollock (personal communication). These beds were penetrated by drill cores as part of the Living Filter site characterization efforts. Fine-grained shale, siltstone, and silty dolomite beds are present at various depths within the Upper Sandy Dolomite Member. These greenish beds were

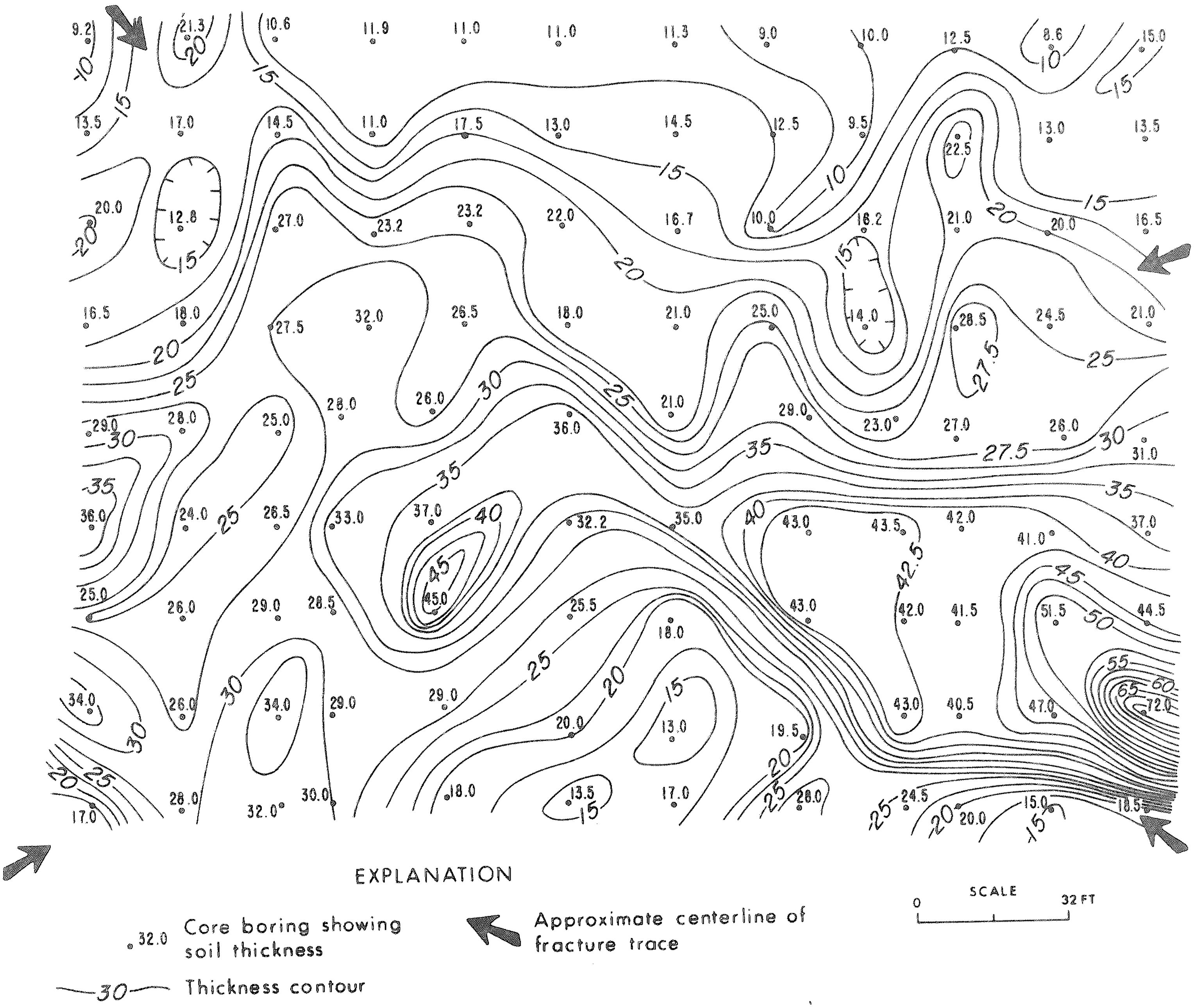


Figure 15. Thickness of residual soil developed below a fracture trace intersection in the Nittany Dolomite, East Halls Cafeteria, Penn State University.

described as being composed mainly of clay and quartz based on a binocular level ot inspection. Later, Pollock showed them to be rich in feldspar, the parent material for kaolinite observed in the Barrens. The problem was how to form the thick kaolinite clay deposits (Stop 2) given the predominantly dolomitic and quartz sand nature of the Upper and Lower Sandy Dolomite Members of the Gatesburg. R. Pollock reports that some of these clay deposits are nearly pure kaolinite and highly refractory. Following carbonate dissolution and chemical alteration of feldspar, quartz had to be leached from the clay. This would be possible under a prolonged tropical to subtropical climate where silica is more soluable than at present. E. G. Williams and R. Pollock both identified the mineral bauxite associated with these kaolinite deposits which offers further evidence for a tropical to subtropical paleoclimate in Nittany Valley.

Evidence for Late Paleocene through Early Eocene bauxitization in the southeastern states and Arkansas under a humid tropical climate helps to explain the sparse occurrences of bauxite in the Barrens which is located more than 500 miles to the north. Barrens deposits may have formed under a more temperate zonal climate that extended to the north into Pennsylvania.

Pollock offered an explanation of how kaolinite deposits might have become so thick in the Barrens regions lacking outcrops or cores showing thick feldspathic sands in the underlying Lower Sandy Dolomite Member. To the west of the Birmingham thrust fault, beds are vertical to near vertical or steeply dipping (west of Stop 2). Selective weathering along bedding planes and along beds with high feldspar content would produce long narrow beds of kaolinite. These deposits would accumulate as vertical weathering continued. Thick, near vertical beds of kaolinite would result that display relict bedding. Where original bedrock had more gentle dips, kaolinite deposits would be let down over a wider area and subjected to differential erosion. Alternatively, thick irregular beds of feldspathic sand may have been present in the Gatesburg that have been altered to kaolinite deposits observed although silty and sandy dolomite beds are known to occur in the Lower Sandy Member, no irregular feldspathic sandstone bodies have been observed to date that might account for irregular occurrences of Barren's kaolinite.

It is unlikely that Barrens' kaolinite deposits are of primary marine origin. Marine environments are alkaline, there is no leaching, and water contains a good deal of dissolved calcium. According to Grim (1953) these environmental conditions favor the formation of montmorillonite, illite or chlorite clay minerals rather than kaolinite. Millot (1942) indicated that Ca^{++} tends to block the formation of kaolinite. To produce kaolinite from a dolomite parent material, carbonate minerals must first be leached. The pH of the geochemical environment had to be in the 5 to 9 range where silica becomes more soluable in order to remove quartz.

Grim (1953) indicates that primary kaolinite, if deposited in a marine environment, is likely to persist because diagenetic alteration of kaolinite is slow. Relict bedding observed at Stop 2 is more indicative of a weathering alteration mechanism of kaolinite formation rather than primary deposition. The pod-like character of Barrens kaolinite deposits also does not support a primary deposition model for this kaolinite. Dark gray dolomite units in the Gatesburg Formation are interpreted as having been deposited in an off-shore environment. Individual beds may be traced for a considerable distance when compared to beach sands and lagoonal facies. Clay deposited in an offshore marine environment most-likely would produce sheet-like deposits rather than the discontinuous, lense-like deposits observed. Sands and shales in the Gatesburg on the other hand, are lense-like and are the most probable source of feldspars that gave rise to these kaolinite deposits.

There is no evidence to support a hydrothermal origin for the Barrens kaolinite as are sometimes found as an aureole around metalliferous deposits. Elsewhere, a supergene origin for kaolinite has been found associated with metalliferous sulfide ore bodies. Grim (1953) reports that such clays are developed during the downward movement of acidic water produced by the oxidation of the sulfides. Lead and zinc have been mined in the Milesburg Gap area near Bellefonte (Stop 7) and near Birmingham, Pennsylvania, but such deposits if responsible for Barrens kaolinite, have long since been eroded.

A soil weathering mechanism offers the best explanation for the origin of Barrens kaolinite and immediately adjacent beds of iron ore. Under long periods of continued weathering in a hot wet environment, organic acids and a neutral or slightly alkaline environment ideal for lateritic alteration might develop that allow for magnesium removal, silica to be carried away, and iron and alumina to be concentrated at or near the surface. In weathering calcareous sediments, there is substantially no alteration of silicates until the carbonate is completely broken down and calcium removed from the environment by deep leaching. After calcium is removed, development of a zone of aluminum and iron concentration are possible (Grim, 1953).

Alluvial Soils

Baxter (1983) mapped two prominent, longitudinally continuous terrace levels along Beech, Fishing and Bald Eagle Creeks and West Branch of the Susquehanna River Valley. Flood plain alluvium comprises the lower level and intermediate terrace alluvium, the higher level. He also mapped a discontinuous terrace alluvium between and above these two main levels.

Depending upon where the pre-Wisconsinan glacial boarder is mapped below Lock Haven (figure 16) several origins are possible for these terrace remnants. If ice

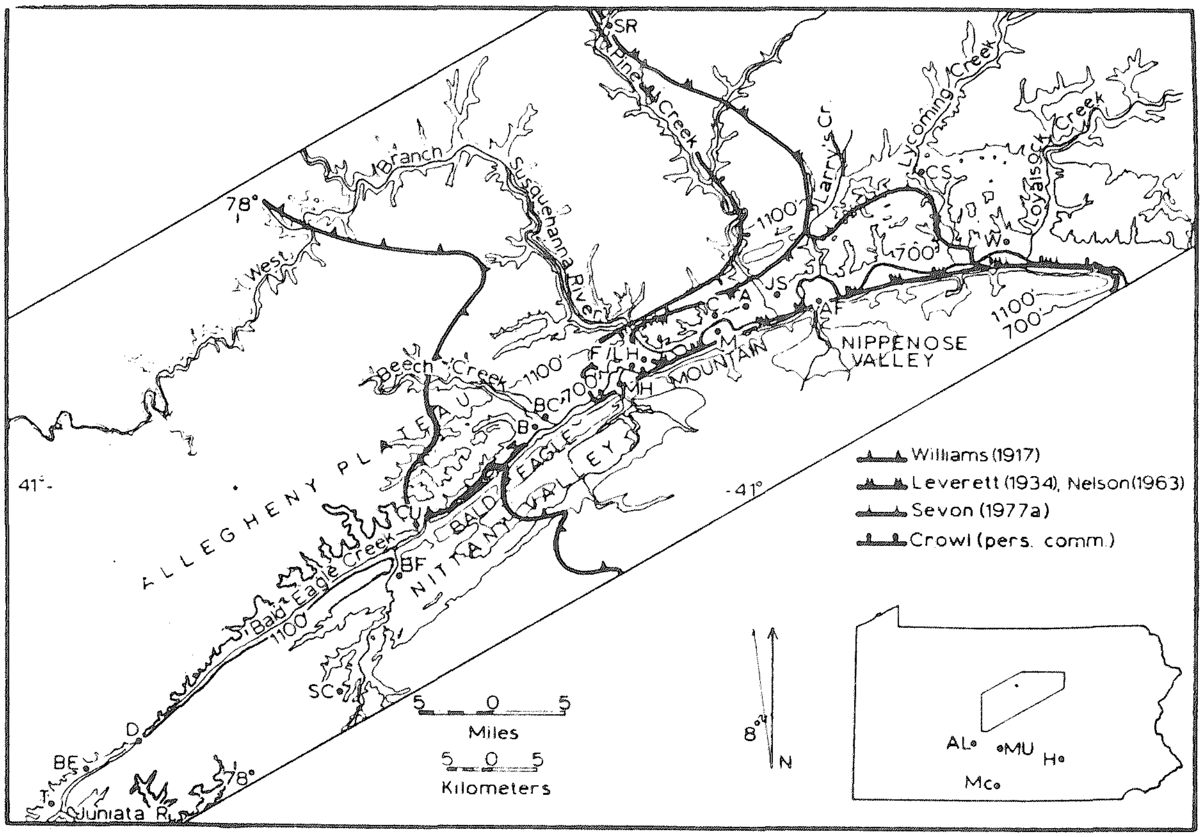


Figure 16. Proposed positions for a pre-Wisconsinan glacial border within the West Branch and Bald Eagle Creeks near Lock Haven (from Baxter, 1983). The 1,100 ft contour shows extent of Lake Lesley (Williams, 1895, 1917, 1920). 700 ft contour shows extent of the lower elevation proglacial lake (Leverett, 1934; Bucek, 1975, 1980; Sevon, 1977a). Location abbreviations are: A-Avis; AF-Antes Fort; AL-Altoona; B-Blanchard; BC-Beech Creek; BE-Bald Eagle; BF-Bellefonte; C-Charlton; CS-Cogan Station; CU-Curtin; D-Dix; F/LH-Flemington/Lock Haven; H-Harrisburg; JS-Jersey Shore; M-McElhattan; MC-Mercersburg; MH-Mill Hall; MU-Mount Union; SC-State College; SR-Slate Run; W-Williamsport.

stood near Jersey Shore, then alluvial terraces were colluviated producing diamictons that were graded to a temporary base level in the West Branch. However, Baxter (1983) rejects this "nonglacial" model as well as a model that proposed a glacial advance up Bald Eagle Valley to Blanchard to explain extensive diamicton deposits to this point. This would require the presence of a proglacial lake within Bald Eagle Valley controlled by a spillway elevation of approximately 1110 feet above sea level at Dix (8 miles west of Stop 1) as proposed by Williams (1917) or a small or short lived lake if meltwater followed the present valley through subglacial and englacial channels in the ice marginal to the end of Bald Eagle Mountain (Bucek, 1975; Baxter, 1983). Baxter (1983) rejects this more extensive glacial model on the best available data. Rather he presents evidence that a pre-Wisconsinan glacier advanced to the vicinity near Lock Haven (Figure 16) and solves the proglacial lake problem by (1) allowing the upper West Branch of the Susquehanna River, Bald Eagle Creek, and Fishing Creek to drain subglacially and englacially around the east end of Bald Eagle Mountain for a distance of 40 miles or (2) the streams to dam for only a short time to preclude significant accumulation of lacustrine deposits within Nittany and Bald Eagle Valleys to an elevation of approximately 1110 feet above sea level. This elevation would allow a lake to extend nearly to State College.

A lower lake level would help to some extent because no lacustrine sediments are known within Nittany Valley to this or even lower levels. The col at Dix is rather narrow but the col could contain Wisconsinan colluvial fill derived from Bald Eagle Mountain. A 10 to 30 foot thickness of colluvium is possible which would result in a 1100 to 1080 foot lake level.

Baxter (1983) uses the retreating ice to help aggrade these valleys with fluvial sediments to the reconstructed terrace longitudinal profiles. He regards the intermediate and upper terraces in Fishing Creek (below Lamar) as old alluvium because this valley was isolated from continental glaciation by Bald Eagle Mountain. He states that these terraces are found only in terrace position along these stream, their reconstructed longitudinal terrace profiles are smooth and parallel to the present stream profile and Beech Creek and Bald Eagle Creek terraces contain clasts derived from the Allegheny Plateau and Bald Eagle Mountain.

Baxter (1983) postulates that the intermediate terrace level resulted from the rapid influx of colluvium into these drainages as an Illinoinan glacier retreated from its position at Lock Haven. Incision followed shortly after to produce the intermediate terraces.

Deep weathering followed during the Sangamonian and this was followed by Early Wisconsinan glaciation to the vicinity of Loyalsock Creek near Williamsport (Bucek, 1975). This Early Wisconsinan ice dammed the West Branch at Williamsport to produce proglacial Lake Lesley II which was less extensive than Lake Lesley I proposed by Williams earlier. According to Baxter (1983), glacial outwash near Williamsport allowed the aggradation of the Susquehanna River and the development of the lower alluvial terrace levels at Lock Haven, along Fishing Creek and near Beaver Creek.

The terrace noted at Houserville (Stop 5) most likely correlates with the intermediate terraces of Baxter (1983) lower in the Bald Eagle and Susquehanna drainage system judging from its well developed weathering profile. This terrace does not owe its origin to younger fan alluvium associated with Late Wisconsinan glaciation because no such fans impinge on Spring Creek between Houserville and Milesburg Gap where Spring Creek joins Bald Eagle Creek. Its mapped extent is shown in figure 17.

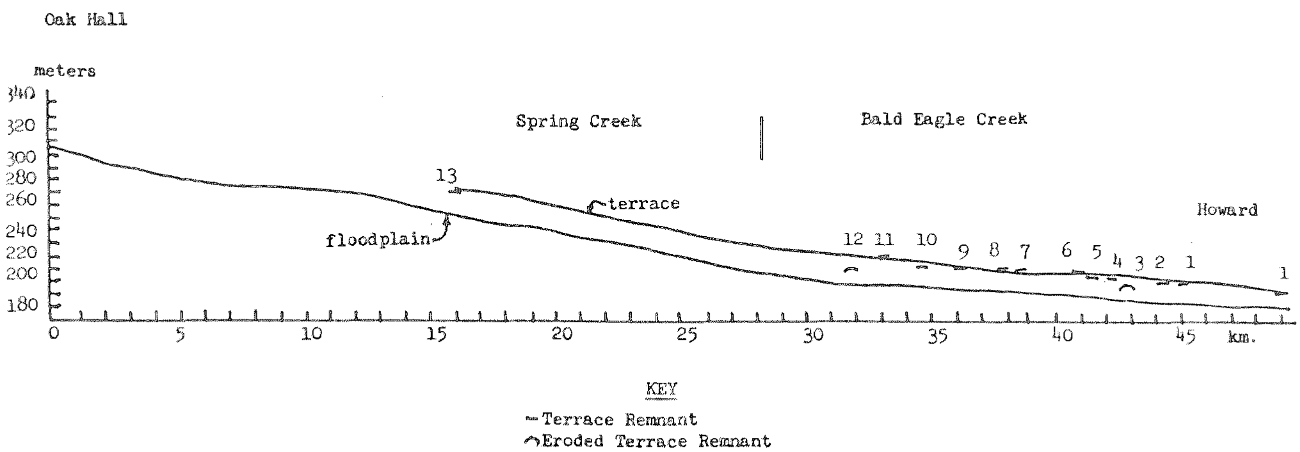


Figure 17. Floodplain and terrace longitudinal profiles along Spring and Bald Eagle Creeks to Howard, Pennsylvania (from Kashatus, 1984).

The presence of a buried soil profile below Spring Creek alluvium is of interest. This soil underlies recent alluvium by a depth of 4 to 8 feet. It is not known if this buried surface correlates with charcoal dated nearby by Bilzi and Ciolkoz (1977) at 200 Y.B.P. A problem remains; when was this soil developed?

Sewer main construction encountered bedrock at depths of 6 to 12 feet near the confluence of Spring and Slab Cabin Creeks. No deep bedrock channel stage was recognized below this alluvium that would allow this buried surface to be elevated for a long enough period to produce the soil profile observed. How might such a soil profile develop below the active flood plain level? Spring Creek is influent near Stop 5. The water table stands 12 feet below the level of Spring Creek at present but it is not clear how this might have allowed the development of a soil profile in a flood prone setting.

A cattle underpass was constructed under the State College Bypass just across Spring Creek (Stop 5). Trench walls at this site revealed the presence of cyclic sediments resembling varves to a depth of more than 6 feet. Is this the long sought after evidence for Lake Lesley I? Unfortunately, this deposit may have been produced by iron ore washing activities during the last century. Extensive excavations for iron ore have been covered over just beyond the bypass and below the Mountainview Hospital. These finely laminated deposits noted on Spring Creek flood plain by the cattle underpass may be related to orewashing activities rather than a proglacial lake. This surface was flooded during Agnes (June, 1972) but there is no chance that these cyclic sediments record a past frequent flooding history similar to Agnes. Biological processes destroyed bedding in Agnes overbank sediments shortly after they were deposited. These cyclic deposits observed had to have been protected from such biologic activity.

The buried soil observed may be of Early Wisconsinan age and does not appear to have been caused by iron mining activities. The soil underlies a distinct flood plain surface and alluvium, not man-made land.

No evidence has been found in the region to support the suggestion that significant quantities of the overburden was deposited by rivers or a marine transgression sometime in the distant past. The Cretaceous outlier located near Chambersburg, Pennsylvania appears to be a unique occurrence. It was an isolated lignite deposit not part of a marine transgression.

Colluvial Soils

Mountain slope colluvium and alluvial fan complexes are well developed opposite water gaps and along the lower slopes of Bald Eagle, Nittany and Tussey Mountains (Figure 18). These deposits overlie residual soils and are distinctive. They tend to thin out in an irregular manner as they are traced out onto residual soils of the valley upland. They may thicken significantly where they were deposited in sink and swallow holes, and closed surface depressions reflecting paleokarst landforms formed during interglacials when processes of colluviation were largely inactive.

In the Barrens region, for example, residual soils have been reworked and transported up to a mile or more along intermittent drainage ways. Distinctive pebbles, cobbles and boulders of chert, quartzitic sandstone, and oolitic chert may be traced down slope across younger bedrock units that are located 2,000 feet or more higher in the stratigraphic section. These deposits were worked for iron ore in the early days having been excavated to depths of 20 to 30 or more feet in open pits and shafts. Such deposits were termed "wash ore" and are lag concentrates of iron oxides, sandstone, chert and other impurities. These deposits show a sharp contact with underlying residual soils, have distinctive color and a gravelly texture when compared with underlying residual soils.

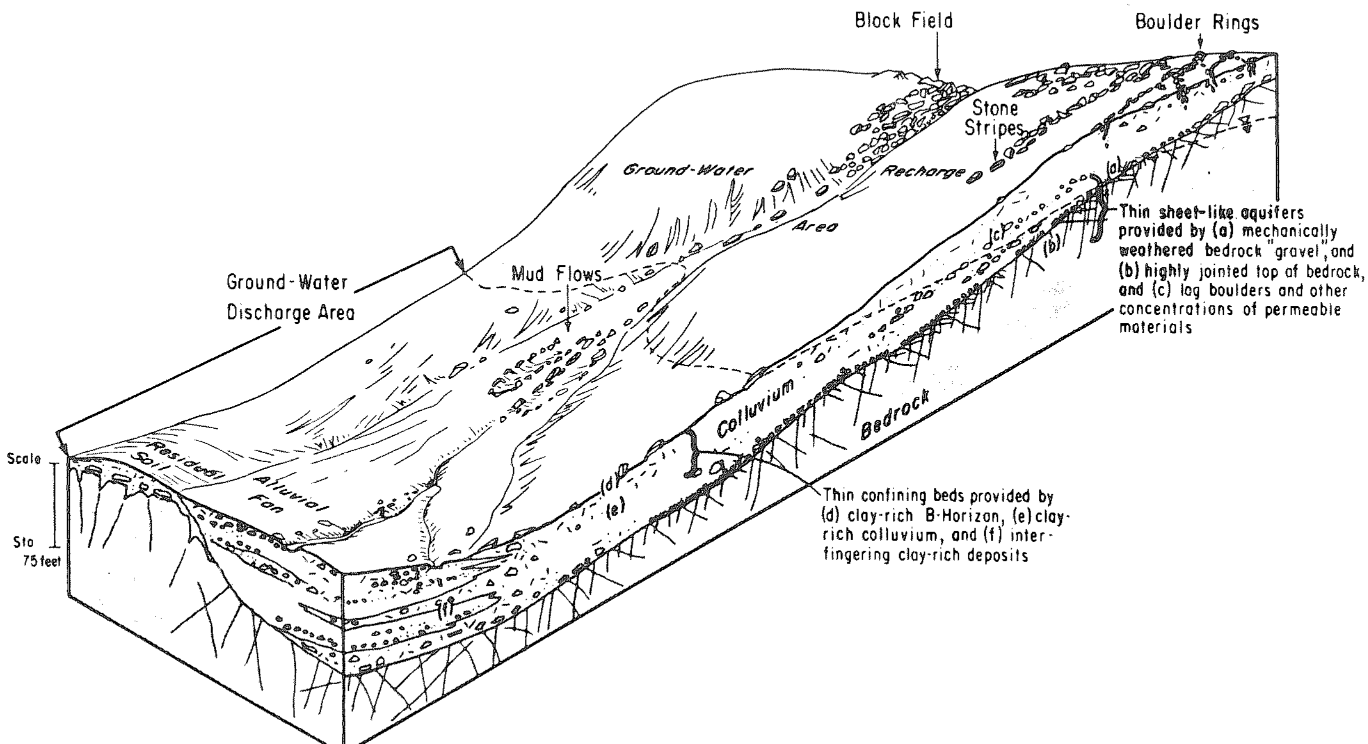


Figure 18. Nature of colluvium and colluvial-alluvial fans and other periglacial landforms in the Ridge and Valley (from Parizek, 1971).

Where the Gatesburg Formation stands in topographic relief above adjacent bedrock units, colluvium may appear as a blanket of surficial sediments of variable thickness (2 to 8 or more feet). In an area in Huntingdon County, southeast of Stop 2, characteristic swell and swail topography post dates this colluvium. Differential dissolution of bedrock has resulted in the development of closed surface depressions with 5 to 10 or more feet of relief. These depressions may be 50 or more feet wide and elongate parallel to regional joint sets. The surficial landform is younger than the thin vaneer of colluvium indicating that regional carbonate rock dissolution has been appreciable since the colluvium was deposited.

Soil characterization studies by Penn State Agronomists are helpful in distinguishing colluvium. Soils of the Meckesville and Laidig groups (Table 1) have a large component of sandstone material in their colluvium; hence, have medium to coarse textures. Soils of the Murrill group are associated with the Laidig and Meckesville groups but are located farther downslope where colluvium extends out over limestone bedrock. The Murrill soils have a significant quantity of limestone residuum mixed into them which accounts for their higher clay contents and fewer rock fragments (Ciolkosz et al., 1979).

The Mertz groups of soils is associated with low ridges and mantles limestone and shale bedrock. They are generally located on the upper side slopes of these ridges and have been less influenced by underlying limestone and shale than the Kreamer and Evendale soils located farther downslope. This makes the Mertz soil slightly coarser textured than the Kreamer and Evendale soils.

Ciolkosz et al. (1979) suggest that the colluvial material was moved and deposited in thinner sheets or flows on the lower ridges than on the higher major ridges. All contain rock fragments derived from up slope bedrock and older colluvium. The fragments are sandstone with some shale in the soils of the Laidig, Meckesville, and Murrill groups and chert in the soils of the Mertz group. The authors characterize rock fragments as follows:

(1) Fragments of chert are usually < 7.5 cm in diameter and equidimensional while sandstone fragments are flat and variable in size,
(2) high concentrations of large sandstone fragments may form pavements on the surfaces of the soils of the Laidig and Meckesville groups, and
(3) the soils of the Murrill group generally have fewer coarse fragments than soils of the Laidig and Meckesville group.

Ciolkosz et al. (1979) report that colluvial soils show a general trend of less thorough leaching from the well-drained to the more poorly drained members of the colluvial soil sequences. This trend has been noted for other soils with fragipans (Peterson et al., 1970) as well as for soils without fragipans (Ranney et al., 1974) Ranney et al. (1974) points out that percent base saturation is a more sensitive indicator of leaching than pH.

Soils of the Laidig and Meckesville groups and the Clarksburg soil have fragipans, though the Murrill soil and Mertz group soils do not (Ciolkosz et al., 1979). The fragipan horizon in the Meckesville and Laidig soils are deeper in the profile than their more poorly drained associates. Ciolkosz et al. (1971) indicates that there is a general trend of increasing fragipan development progressing from well-drained to somewhat poorly drained soils. Fragipans display a firmness, brittleness, and mottling reflecting poor drainage. These soils also show a coarse prismatic structure and higher bulk density than other soils.

Mertz, Kreamer, Evendale, and Murrill soils lack a fragipan, and the Clarksburg, Laidig and Buchanan soils have fragipans. The lack of a fragipan appears to be related to their high clay content hence fine texture, and limestone influence (Ciolkosz et al., 1973).

Although colluvial soils show major textural variations both vertically and laterally, argillic horizon development has altered the texture of the upper meter of the soil by depleting the A and enriching the B horizon with clay (Ciolkosz et al., 1979). They indicate that this clay migration has not obscured the original textural heterogeneity of the parent material. For example, argillic horizon development is weak to moderate in soils of the Laidig and Meckesville groups and moderate in soils of the Murrill and Mertz groups. The alluviation of clay (< 0.002 mm) from the upper horizons and deposition in the B horizon, presence of clay skins on ped faces and in soil pores all indicate that these colluvial slopes have remained static for a prolonged period and allowed the development of distinct soil profiles.

Further evidence for soil profile development in the distinctive relationship of decreasing expandable-layer silicates (mainly vermiculite) and increasing illite content with depth. Johnson et al. (1963) indicate that this is a useful weathering index for Pennsylvania soils. Illite tends to be converted to illite weathering products or vermiculite and other expandable silicates. The ratio of expandable to nonexpandable materials is used as a weathering index. It is most intense near the surface and decreases to a relatively constant level in the middle and lower parts of the profile. Weathering ratio data also indicate that as soils become more poorly drained, there is a decrease intensity of weathering paralleling the trend of higher base saturation resulting from less thorough leaching.

Time is required for fragipan formation, significant leaching, and clay mineral weathering. Soils with argillic horizon are found on stable landscapes that are known to date back to many thousands of year (Soil Survey Staff, 1975). Ciolkosz et al. (1979) indicate that fragipans in soils also indicate landscape stability. Nearly all transported soils developed on glacial drift of suitable texture have fragipans where as they are absent in recent floodplain soils. Bilzi and Ciolkosz (1977) studied soil development in four floodplain soils ranging in age from 200 to 2000 years BP including the Nolin soil on Spring Creek near Houserville, Pennsylvania (Stop 5).

Soil formed in these alluvial deposits lack a fragipan or an argillic horizon. Both time and landscape stability are required for these to be present. Leaching of soils extensively enough to form ultisoils also requires time and stability as does the observed clay mineral weathering index properties noted. The regular decrease of the lwp: Illite ratio with depth, and its close similarity to that noted for Wisconsinan age till soils of similar drainage and texture indicates that these soils probably date back to Wisconsinan time (Ciolkosz et al., 1979).

Aside from these lines of evidence, other observations support evidence for stability of colluvial slopes and alluvial-colluvial fan complexes. The largest trees show an upright growth position as do stumps related to past timber cutting activities. Railroad and road cuts have remained stable for 100 or more years with some notable exceptions where cuts were placed above shale within groundwater discharge areas (Parizek, 1971). Building, railroads and other structures are rather stable despite the fact that often, no special foundation design precautions were taken. Exposed surfaces of boulders derived from the Oswego and Tuscarora Formations show a weathered, sandy, dull surface character indicating that they have remained upright for a prolonged period compared with their fresher, smoother, and brighter colored undersides. Lichens and lichen rings etched into sandstone surfaces also are evidence of slope stability.

The association of colluvium with other periglacial features such as patterned ground (Walters, 1978; Clark, 1968; Parizek, 1971) boulder fields, grezes litees and involutions (Marchand, 1978) flow lobes and store stripes (Parizek, 1971) and the similarity in soil development between these soils of colluvial deposits and

Wisconsinan glacial till deposits also indicates Wisconsinan periglacial movement and deposition (Parizek, 1971; Marchand et al., 1978; Ciolkosz et al., 1979).

Most likely the last active period of colluviation occurred during the Woodfordian because soils have an appearance not unlike soils developed on Woodfordian tills. In glaciated regions to the northeast of Nittany Valley, Woodfordian till was colluviated and has since been stable long enough to allow profile development.

At least one other major episode of colluvium development is commonly recognized. Where exposures are deep enough and colluvial deposits are well developed, a truncated paleosol of variable thickness sometimes can be observed. This soil has a distinct "Allenwood character". It contains an abundance of clay films and coating on pebbles and cobbles, abundant iron enrichment as coatings on rock fragments and mixed with matrix fines. Distinct weathering rhins on cobbles and pebbles also are noted. The brightness of color and overall appearance of these buried soil deposits strongly suggests that they are truncated paleosols of at least Sangamonian age if not Altonian age. These older colluvial deposits often are thicker than the younger Wisconsinan colluvium that covers them suggesting that residual soil deposits developed before Illinoian glaciation probably were thicker than the soils available for movement during the Wisconsinan or that the Illinoian periglacial climate was more severe and persistant. Illinoian glaciation was more extensive in northeastern Pennsylvania than Wisconsinan glaciation. It would be reasonable to expect more extensive development of colluvial deposits in proglacial settings such as Nittany Valley because of its closer proximity to the Illinoian glacial border than to the Wisconsinan borders.

Older colluvium often shows a sharp contact with younger, dull colluvium that overlies and truncates the paleosol. Elswhere, bright "Allenwood type" soil deposits are included in overlying younger colluvium either as iron coated rock fragments, and rock fragments with weathering rhins (Stop 7). Masses of weathered colluvium also can be contorted or balled up into younger colluvium suggesting that mass movement of soil involved structural development and flow as well as downslope creep of individual rock particles.

Where colluvium overlies the Reedsville and other shales, shale chip gravel deposits sometimes are well developed. They have all the appearances of grezes litees but are usually overlain by more typical sand, silt, clay boulder enriched colluvium. Greze litee requires a periglacial climatic regime. It is interesting to speculate on the timing of these deposits. Greze litee would have to be derived from shale outcrops exposed upslope before more coarse textured sandstone float and matrix fines moved downslope and buried these shale chip deposits. This suggests that the greze litee environment may occur slightly earlier than the environment favoring colluvial development or time was required for colluvium to overwhelm shale exposures that may have stood slightly higher in elevation than adjacent shale surfaces.

The irregular thickness of colluvium noted below water gaps and mountain slopes and below the Salona-Coburn or Reedsville contact, or shales stratigraphically below but upslope from the Old Port Formation indicate that sink and swallow holes existed in these settings before colluvium development. Depressions 20 to 150 feet or more deep and of variable size revealed by water well records had to be filled with colluvium and alluvial deposits before the transport slope was established that allowed for continued downslope movement of colluvium. Post-glacial dissolution of limestone and inwash and collapse of colluvium into paleokarst channels and voids has occurred since the retreat of the last glacier. Sink- and swallow-hole development is extensive in some cases despite the relative brief period that has elapsed since the retreat of the last glacier 18,000 to 12,000 yrs BP. Some existing sink and swallow holes are 30 to 50 or more feet deep and up to 100 to 300 feet or more in diameter suggesting that some paleokarst sink filling debris must have been flushed into underlying conduits rather than having formed soley as a result of post-Woodfordian limestone dissolution.

Boulder fields and thick colluvium also have been observed draped into karst depressions high on the western flank of Bald Eagle Mountain where the surface watershed area is limited. The McAlevy's Fort-Port Matilda lineament cuts across this portion of the mountain. Sinkholes formed in this setting in post-Woodfordian time most likely were formed by lateral flow of groundwater moving parallel to stratigraphic strike from a larger watershed rather than just from immediate upslope.

Aside from sinkhole collapse, many colluvial-alluvial fan complexes opposite water gaps and blanket type deposits between gaps show gullie development and only minor evidence of stream erosion. Although some of this erosion post dates clear cutting of mountain slopes during the peak of iron mining and early attempts at cultivation of steep mountain slopes, most of this erosion appears to predate European settlement.

Evolution of Nittany Valley

A modified version of Davis' (1889) model for the evolution of Appalachian topography illustrating Nittany Valley is presented in figure 19 (Gardner, 1983). Ciciarelli (1971) elaborated on certain aspects of the initial stages of the model in Sugar Valley. According to Gardner (1983), Ciciarelli's model can be segmented into four convenient stages. The model starts with an initial surface of Tuscarora Sandstone that is folded into anticlines and synclines where anticlines are structurally and topographically the highest point on the land surface. First order consequent streams (streams without tributaries) develop along the flanks of anticlines and drain into larger, consequent streams in synclines. Synclines act as focal points for collection of drainage. Consequent synclinal streams drain to the northeast, down the structural plunge. The Tuscarora is breached by first order streams at the structural culmination along the crest of each anticline. Ciciarelli (1971) has shown that breaching initially occurs at the structural culmination as a result of increased fracture density in that area. The Tuscarora is mechanically weakened and more susceptible to erosional processes. Cliffs form in the Tuscarora and erode headward, exposing the less resistant Juniata. Small drainage basins with subsequent streams develop as the valleys expand in less resistant rock.

In the second stage individual basins are separated by cols of Tuscarora. Drainage divides migrate headward and basins coalesce. As downcutting continues, the Oswego is exposed along the topographically high anticlinal crest and the characteristic double inner and outer ridge takes shape. During the third stage the drainage line which has an exist gap at the lowest base level (furthest down the plunge of each anticline) captures drainage from other subsequent basins. As the Tuscarora and Oswego cliffs erode headward, anticlinal valleys continually gain surface area at the expense of consistantly shrinking synclines. Cambro-Ordovician carbonates are exposed in the anticlinal core. Since carbonates are less resistant than the sandstones, a process of topographic inversion takes place. The limestone

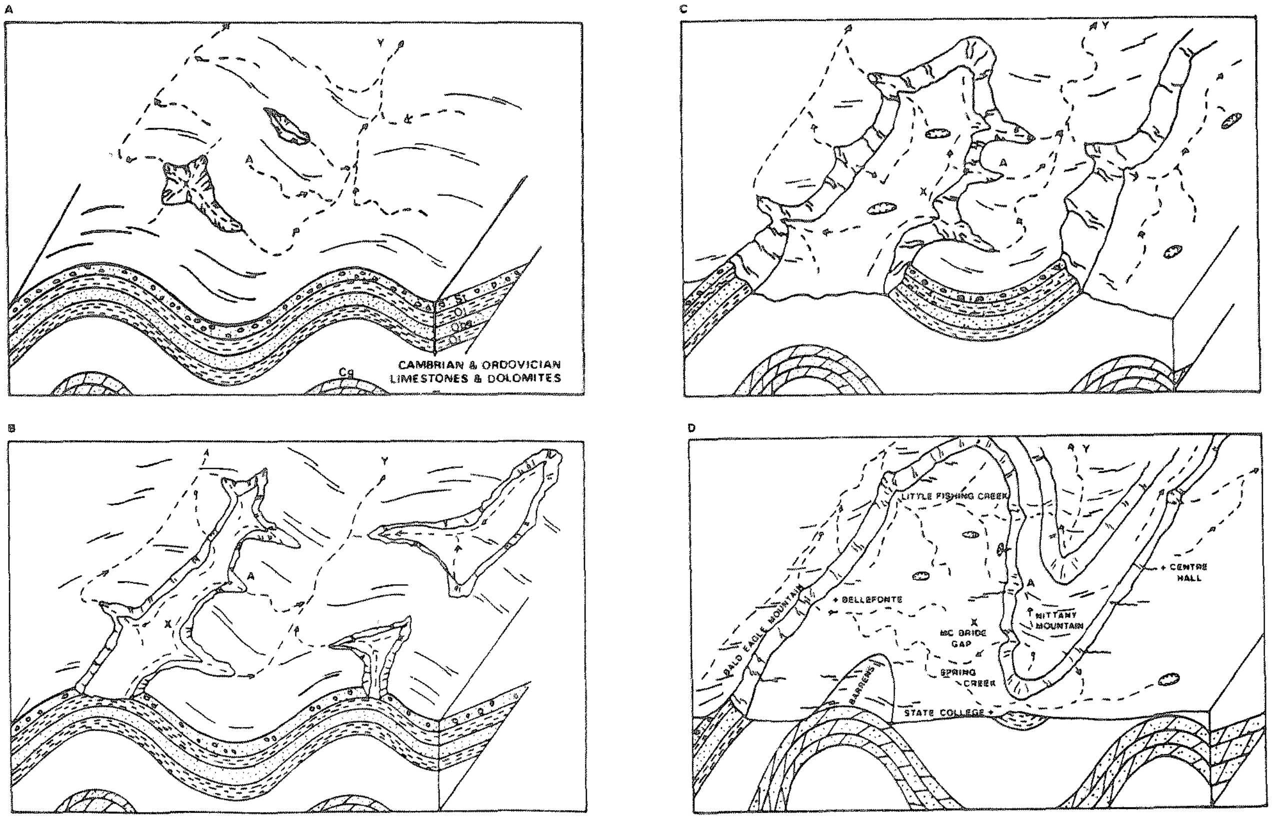


Figure 19. Four stage (a, b, c and d) model for the geomorphic evidence of Nittany Valley. Y is a major synclinal, consequent stream. X is a developing anticlinal subsequent stream. A is a fixed reference point showing the capture of the synclinal stream in part C (from Gardner, 1983).

surface is lowered faster than the adjacent sandstones: the originally higher anticlines become topographic lows while the lower synclines become topographic highs. With continued topographic inversion drainage is diverted from the synclinal axes and an integrated drainage pattern is established in the newly formed anticlinal valley.

During stage four, major drainage lines in Nittany Valley assume their modern configuration. The present surface takes shape and topographic inversion reaches its culmination.

Gardner (1983) points out an interesting complication to this model arising from the fact that downcutting may not have been continuous during all stages of topographic development. According to Davis, the ideal cycle of erosion and peneplaination begins with rapid uplift of a land mass. It is followed by a long period of tectonic quiesence. During the period of stability the landscape progresses through a sequence of stages; youth, maturity, and old age. As the region under consideration passes from one stage to another, its characteristic features gradually change as once mountainous areas have been worn down by erosion. In the end, at old age a flat, featureless plain, or peneplain, develops at the regional base level.

It has been suggested that three major episodes of peneplaination are preserved in the Appalachian Mountains. From oldest to youngest, they are the Schooley, Harrisburg and Sommerville Surfaces. The evolution of those peneplains from initial Appalachian folds is depicted in Figure 20. The Schooley Surface is reported to be either Cretaceous (Davis, 1889) or early Tertiary (Johnson, 1931) in age, but data are equivocal. It is one of the most complete cycles, effectively beveling the folded sedimentary rocks in Pennsylvania and adjoining areas. One of the most striking attributes of the physiography of the Nittany Valley area, the accordant summits of the Tuscarora Ridges, is thought to have resulted from Schooley peneplaination. Superposition of major drainage lines from the Schooley Surface has been suggested as a mechanism for transverse Appalachian drainage, where streams cut across the stratigraphic and structural strike.

After Schooley peneplaination the area was subjected to renewed uplift and stream downcutting. Valleys were opened in less resistant rock types. Another pose in downcutting resulted in the formation of the Harrisburg peneplain in middle Tertiary time. Thus, the Harrisburg Surface has been termed a partial peneplain. The wide expanse of Nittany, Penns, and adjacent valleys marks the Harrisburg surface. Development of the Sommerville Surface is marked by incised valleys of major drainage lines in Nittany Valley. Further southeast, it is more fully developed on carbonate rocks of the Great Valley. Incision is thought to have occurred in latest Tertiary or Quarternary time.

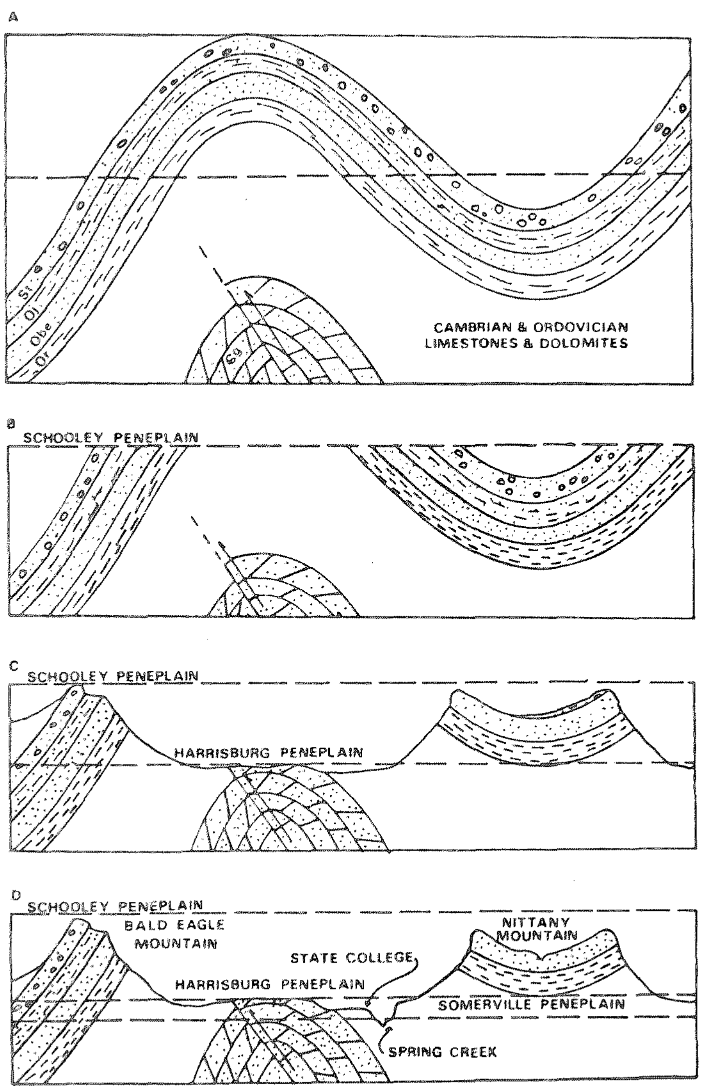


Figure 20. Evolution of the peneplain surfaces of Davis in the Nittany Valley (from T. Gardner, 1983).

Hack (1960) proposed an alternative model for the evolution of Appalachian topography. His theory of dynamic equlibrium maintains that "... the landscape and the processes molding it are considered a part of an open system in a steady state of balance in which every slope and every form are adjusted to every other. Changes in topographic form take place as equlibrium conditions change, but it is not necessary to assume that the kind of evolutionary changes envisioned by Davis ever occur." Differences in topography are thus explainable in terms of differences in the erodibility of bedrock (Ashley, 1935). Using this model, Gardner (1983) shows how the multi-level landscape of Nittany Valley is attributed to differences in bedrock erodibility rather than peneplaination. The Tuscarora and Oswego stand as ridges above the valley floor because they are more resistant. Furthermore, accordant summits of the Tuscarora ridges are a result of the nearly uniform resistance and thickness of that formation. Water gaps are located along zones of structural weakness.

Using the inversion model concept shown in Figure 20, it is interesting to speculate on how the more recent details of landscape evolution might have taken shape through time. If similar relative relief is maintained between more resistant sandstone and shale beds and carbonate rocks to what is observed at present, then the present erosional surface can be incrementally projected upward in space with some degree of confidence at least until the synclinal roots of the next overlying, more resistant clastic sandstone unit is encountered. This exercise is revealing because it shows how drainage divides developed along resistant ridges migrated downdip as the landsurface was lowered. The development of valleys above beds of low resistance and the characteristic trellis pattern emerges. However, a mechanism is required to initiate and maintain transverse valleys, the same problem faced by earlier workers. The answer to this problem was made possible with satellite images and plate tectonics theory.

Some information of carbonate rock removal can be obtained by examination of insoluble residues that remain as residual soils. Interfluves between major drainages are good places to search for thick residual soils especially above bedrock that is deeply underdrained. If the infiltration rate is rapid enough, overland flow and surface erosion are minimized. However, a fine-textured karst is more favorable than a coarse-textured karst because internal soil erosion will be less for the former case. A silty to sandy carbonate bedrock has an added advantage. Clay, silt, and sand tend to bulkup within solution openings thereby reducing the amount of internal soil erosion. Further, as the sand content increases within carbonate rock, secondary porosity and most likely primary porosity also increases (Parizek et. al., 1971). Beds with greater intergranular permeability and porosity for whatever reason, tend to enhance deep dissolution of carbonate bedrock and differential weathering of its surface. This is apparent by studying cores obtained from the Mines and Upper Sandy Dolomite Members of the Gatesburg Formation. As the sand content increases within dolomite, weathering and cavity development increase. Conduits and cavities are best developed in sandy units, and are rarely observed within purer beds of dolomite located either above or below the water table.

Core samples were selected to represent the lithologic variations noted for the Mines Dolomite and Upper Sandy Dolomite Members of the Gatesburg Formation. These were drilled as part of site characterization efforts for Penn State's Living Filter Project (Parizek, 1963, 1974). Samples were crushed to a workable size and 50 gram samples treated with (1:4) HCl for 24 hours, checked for carbonate and retreated one or more additional times as needed. Final residues were rinsed to remove salts and allowed to stand in water for 24 hours. Samples were oven dried and weighed when dry and cold to obtain the weight percent of the original sample that is insoluble residue (Table 2).

Table 2. Insoluble residue for selected cores taken from the Mines Dolomite and Upper Sandy Dolomite Members, Gatesburg Formation.

Well Number	No. of Samples	Wt. Percent of Insoluble Residue	Depth Internal Sampled and Total Thickness, Ft.
F-1	37	3.79	148.2 to 215 (66.80)
F-2	30	4.46	149 to 204 (55)
F-3	31	3.43	95 to 154 (59)
F-4	21	7.54	167 to 198 (31)
F-5	48	4.86	121 to 183 (62)
F-6	14	19.51	28 to 41 (13)
Totals	181	7.27	286.8 Ft.
Outcrop	16	10.25	Elevation 1066 to 1121 (55 Ft.)

To get an idea of the amount of carbonate rock removed, consider a specific soil column of depth 162 feet to bedrock. Given 7.27% as the average insoluble residue, a bulk density of soil of 1.76 g cm^{-3}, and a density of dolomite bedrock of 2.85 g cm^{-3}, simple mass balance considerations demand that roughly 1400 feet of dolomite must be dissolved to produce the 162 feet of residual soil. This is a minimum figure since removal of soil by solution of some of the components or by piping into solution cavities in the bedrock would demand an even larger bedrock thickness.

The rate of dissolution (karst denudation) of carbonate rock terrains in many parts of the world has been an on-going preoccupation of karst geomorphologists (e.g., Smith and Atkinson, 1976; Jennings, 1981; White, 1984). Factors that enter the denudation rate are temperature, precipitation, and carbon dioxide production in the soil (which in turn is a complicated function of soil character and plant cover). At present, Centre County receives about 40 inches of annual rainfall of which about half is lost to evapotranspiration. Using typical present day data in combination with the denudation rate curve given by White (1984) the rate of removal of carbonate rock is calculated to be 30 mm ka^{-1}. Continuing this exercise in the combination of crude but not totally impossible numbers, the time required for the removal of the 1400 feet of carbonate rock represented by the residual soils is 14

Ma. This would place the oldest surface for which there is any record in the residual soils at mid-Miocene, a result which is at least consistant with geomorphic arguments for the ages of the valley erosion surfaces.

One other result comes out of this number-juggling. There is at present only about 300 feet of relief in the valley uplands. Taken as a proportion of the 14 Ma that they valley floor has been deflating, the oldest cave fragments in the valley uplands cannot exceed 3 Ma, a figure in good agreement with the late Pliocene to early Pleistocene ages proposed for many of the older caves in the Appalachian Highlands.

Evidence of the climatic history of the Appalachians is hidden in the coastal plain and continental shelf sedimentary record, in the geochemical character of residual soils, cave fillings, pollen records of bogs, and in deposits and landforms formed under former climatic conditions. Knowledge of past climates is important to the interpretation of past rates of erosion and sedimentation and landform development.

Sevon (1985) reviewed the evidence for climatic change within the Appalachian region from Late Paleozoic to Recent. Highlights are included here as background.

By Permian time the climate gradually changed from temporate during Carboniferous coal deposition to arid (Schwarzback, 1961). The Triassic is believed to have been arid as well. Hay and others (1982) suggest that local topography may have had a significant influence on basin climatic conditions. They postulate that the average elevation of the Appalachian during Triassic time was about 2 km above base level and sea level may have been 60 m lower than at present. This is in agreement with Hallam (1984). Sevon (1985) believes that much of the Appalachian area was at least semi-arid and because of the general proximity of North America to the equator and its considerable distance from the sea at this time.

Sevon (1985) suggests that arid conditions probably extended into Early Jurassic time as the world climatic trends during the period indicate continued warm temperatures and increased moisture. He further states that some of this humidity may have been due to the Atlantic Ocean that was open during much of the Jurassic. He cites the Cost No. B-3 offshore well and geophysical data that indicate up to a 9 km thick pile of Jurassic sediments were eroded from the Appalachians during this time. Marine coals encountered in this same core suggests that at least coastal areas may have been humid by Late Jurassic (Scholle, 1980). According to Scholle (1977) the COST No. B-2 and COST No. B-3 (Scholle, 1980) sediment records indicate that there was a gradual but definite decrease in clastic input to the eastern seaboard throughout the Cretaceous. This suggests a continued change to a more humid climate, increased vegetation and reduced erosion of coarse clastics. The presence of carbonate sediments rated in the COST wells is offered as further evidence that the landscape was lowered and being eroded at a slower rate than during the Jurassic.

Sevon (1985) indicates that by Late Cretaceous, maximum marine transgression had occurred and climatic moderation spread world wide. Vegetation flourished in both polar regions. Pennsylvania may have been subtropical along with much of the rest of the world. Tropical conditions would have been ideal for intense chemical weathering and slow changes in relief. Such conditions would favor the double plaination surfaces proposed by Budel (1982) and coal swamp in the Great Valley (Pierce, 1965). This would be the time of intense chemical weathering in the Piedmont (Cleaves and Costa, 1979).

Sevon (1985) indicates that the Tertiary climate is known mainly from the Eocene flora of the southeastern states along with some correlations with isolated Eocene floras farther north and the presumed requirements for the formation of the secondary mineral deposits associated with the Harrisburg surface. There were fluctuations of cooling and warming during the Paleocene and Eocene, but overall the trend was one of warming to the Middle Eocene and then gradual cooling to the end of the Eocene (Wolfe, 1978). Bauxitization in the southeastern states and Arkansas occurred in Late Paleocene through Early Eocene (Gordon and others, 1958; Overstreet, 1964) in a humid tropical climate. The climate moderated farther north, but warm humid conditions were present as far north as Vermont where deep weathering produced clay deposits and iron and manganese ores in association with the Brandon lignite (Burt, 1928) on the regional equivalent of the Harrsiburg surface. A cooling trend occurred from the Middle Eocene into the Oligocene (Sevon, 1985).

Sevon (1985) comments on the work of Olsson et. al. (1980) who report that a major lowering of sea level occurred in the Early Oligocene which they postulate correlates with development of glacial ice in the Antarctic as well as general worldwide decline in temperature. There were fluctuations in temperature and moisture during the Oligocene, Miocene, and Pliocene, but in general the overall trend was that of climatic cooling which culminated in the Quaternary with Pleistocene continental glaciation (Blackwelder, 1981; Donnelly, 1982).

The Miocene and Pliocene climatic record is less clear for the Atlantic coastal region. Both arid in the southeast and subtropical conditions as far north as New Jersey have been proposed (Sevon, 1985). By the Pleistocene, colder climates prevailed when at least three major glaciations occurred and possibly a fourth. Watts (1979; 1983) indicate that a tundra vegetation existed up to 100 km beyond the ice boarder about 18,000 years ago. This is when most of the recent periglacial landforms and deposits were formed beyond the Woodfordian glacial border.

We are not in disagreement with Sevon's (1985) summary of the most probable age of the valley upland surface in Nittany and adjacent valleys. He indicates that prolonged physical and chemical weathering during a humid subtropical climate that persisted through Late Cretaceous, Paleocene and Early Eocene produced carbonate valley surfaces of low relief bordered by resistant moutain uplands. This he postulates was the time of thick saprolite development, the formation of thick clay deposits throughout the east coast (Burt, 1928; Potter, 1982; Bridge, 1950) and secondary mineral deposits of iron. Such a climate we believe would help account for the limited amounts of bauxite, thick kaolinite clay and iron deposit observed in the Barrens near State College.

Sevon states that the development of the Harrisburg surface may have occurred over a period that may have been as long as 45 m.y. Our own insoluble residue thickness data and estimated rate of carbonate rock dissolution data suggests a conservative age for the valley upland of 14 million years assuming no solution removal or erosional losses of this residum which cannot be the case.

Olsson et. al. (1980) suggestion of an early Oligocene lowering of sea level may have helped to initiate incision of the Susquehanna, Potomac and Delaware Rivers and their tributaries. We are not aware of any evidence for a mid-Tertiary regional uplift in the Appalachian region that may have initiated this incision.

Pre-Sangamonian soil terrace levels just above the modern flood plain level for local streams incised in the valley upland suggest that these valleys were deeply incised by mid-Tertiary time if not during or just before the onset of Pleistocene glaciation about 350,000 years ago.

This review of climatic data for the eastern U.S. is helpful to the present discussion. Subtropical conditions during Late Cretaceous at least through Middle Eocene are indicated by more than one line of evidence; vegetation, soil development, lignite beds and occurrence of mineral deposits. The thick kaolinite clay deposits in the Barrens region most likely reflect such a time of intense chemical weathering.

Concluding Remarks

The mechanism of carbonate rock denudation includes attack at the soil-rock contact as well as internal dissolution within the rock mass itself. This process produces residual soil that is inverted from is normal stratigraphic sequence. Insoluble residue just released is added to the regoliths from the bottom up with the oldest soil being present at the top. Hence, we are looking at the residuum of rock that has long since been dissolved away, often under climatic conditions vastly different than at present. Residual soils documented to be > 100 to 365 feet thick, may be weathering products that have been accumulating since Early to Late Tertiary if not longer. This is suggested by the great thickness of regolith present, the slow rate of carbonate rock dissolution, say about 40 mm/1,000 years, and the fact that some insoluable residue has dissolved or been eroded away during this period.

The presence of bauxite, thick kaolinite clay and iron ore deposits associated with these residual soils suggests a more humid and tropical climate favorable to laterite production. The Sangamon interglacial also has produced a soil profile far more extensively developed when compared with soils developed on Wisconsinan glacial and periglacial deposits but it is unlikely that this weathering interval contributed much to the overall development of thick residual soils observed in the Barrens region. Colluvium with a buried Sangamon paleosol underlie colluvium most likely of Woodfordian age. These inturn overlie thick residual soils along the lower flanks of mountain slopes which predate Illinoian glaciation and the Pleistocene.

These thick residual soils demand favorable conditions for their preservation: (1) remoteness from the coast, (2) separation from ultimate baselevel by a number of local baselevels provided by resistant sandstones, (3) location on interflues between drainages that have maintained their basic position on the landscape for prolonged periods of time, (4) master streams fixed along vertical planes of weakness for prolonged periods, (5) a deep regional water table and (6) karst conditions that favor internal drainage.

Aside from local pyricies, master streams must have occupied nearly their same water gap portions as hundreds if not thousands of feet of carbonate rock have been removed since the breaching of the Nittany anticlinorium.

Lovers of Nittany Valley will be pleased to learn that the Valley has widened through geologic time as surface water divides migrated in the down dip direction, i.e., Bald Eagle Mountain has remained nearly in its present position for a longer period controlled by its near vertical bedrock dips, Nittany Mountain has grown narrower and Tussey Mountain has migrated eastward. Nothing for the distant forseeable geolgoical future should change this condition because carbonate rocks under lie Nittany and Penns Valleys for depths well in excess of several thousand feet or well below sea level. These rocks will be deeply eroded only if the Appalachian Mountains are uplift above present base level.

In the meantime, several mechanism are in operation. The valley upland can be lowered to the local base level controlled by the elevation of sandstones in Milesburg and Curtin Gaps, Jacksonville-Howard Gap, the Gap near Lock Haven, and Water Street. Ultimately this would produce a low relief karst plain, bordered by moutains. In the mean time, Spring and Bald Eagle Creeks, the Juniata River, etc., will continue to downcut as Pleistocene-aged sediments continue to be removed from storage and rapids on the Susquehanna River are slowly eroded away. Continued down cutting of the local drainage network should help to maintain relief between these incised streams, the valley upland and mountain ridges, and a deep water table and internal drainage. Sea level changes resulting from ice cap retreat or renewed glaciation would have little influence on Nittany Valley.

Record of such future events will be recorded in residual and transported soil deposits that are more likely to be preserved within deep karst depressions developed in the top of carbonate bedrock in a manner similar to the present soils that are preserved in the region.

Table 3. Summary of geological events and possible ages

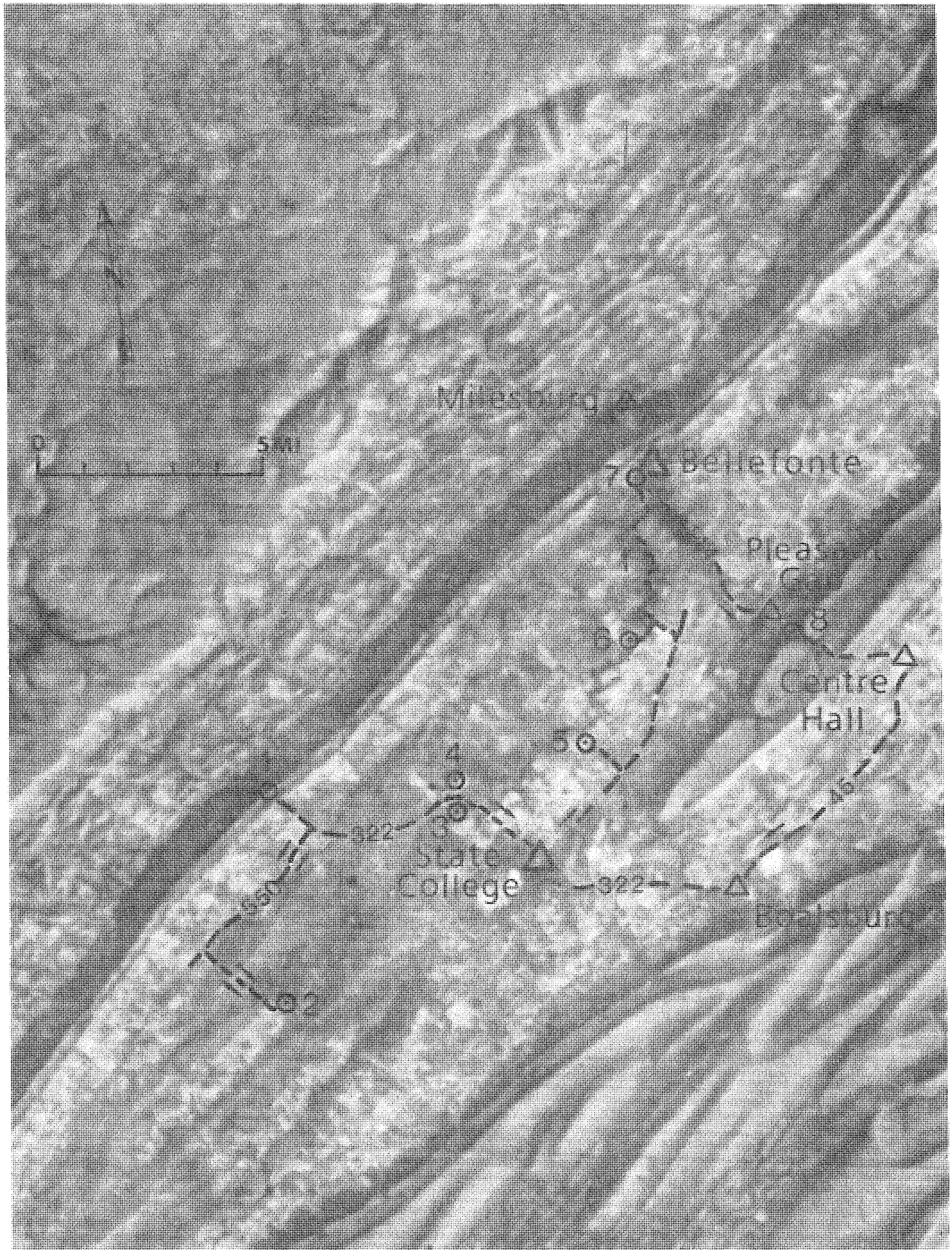
Time and Climate	Features
Holocene (warming trend)	Accelerate sinkhole development due to storm runoff and construction activities 0 to 50 YBP. Accelerate sheet and guille erosion, extensive charcoal production and farming 100 to 220 YBP; cyclic sedimentation due to ore washing. C14 200 to 2,000 YBP on flood plain deposits. Cave travertines $\leqslant$6,000 YBP. Soil development on colluvium 15,000 to present.
Woodfordian Glaciation (cold)	Extensive colluvium and colluvial-alluvial fan complex development on mountain slopes, water gaps and valley upland surfaces; stone stripes, boulder fields, sorted polygons, formed.
(warming trend)	Tundra vegetation to 100 km beyond glacial boarder. Formation of Bear Meadows. Buried soil on Spring Creek(?)
Altonian Glaciation (cold)	Colluvium development (?) and related periglacial landforms and deposits; accelerated erosion.
Sangomonian (warmer than present)	Extensive development of soil on colluvium and alluvium and sapolite ("Allenwood" in character). Terrace development along major drainages; Accelerated development of cave networks graded to terrace levels. Extensive development of speleothems 70,000 to 200,000 YBP.

Table 3. (Continued)

Time and Climate	Features
Illinoian Glaciation (cold)	Extensive colluviation of valleys, mountain slopes and water gaps. Start of development of diamictions. Lake Lesley I; elevation ~ 1110 ft. Cave passage development graded to valley fill.
Early Pleistocene (cooling trend with interglacials)	Deposits and landforms not documented. 800 to 850 ft. elev. cave passages. 900 to 950 ft. elev. cave passages. 1000 to 1100 ft. elev. cave passages. 1200 ft. elev. few remnant cave passages, most likely less than 3 MYBP.
Late Pliocene-Early Pleistocene (cooling trend)	Magnetic reversal in Mammoth Cave, Kentucky.
Mid-Miocene (minimum age)	165 feet of residual soil development in Barrens region, at least 14 MYBP. 365 feet of sapolite even older.
Oliogocene-Pliocene(?) (cooling trend mid-Eocene to Oliogocene)	Incision of master streams; Bald Eagle, Spring and other creeks; start of extensive soil erosion within the Great Valley (?).
Early Oliogocene Late Eocene (cooling trend)	Drop in sea level, ice cap in Antartica.
Early Eocene (warming trend, tropical to subtropical)	Bauxite formation in Ankansas, warm and humid.
Late Paleocene to Early Eucene	
Late Cretaceous (Subtropical to tropical)	Maximum marine transgression; formation of thick sapolite, kaolonite, bauxite, iron ore deposits. Development of Little Juniata River (?); lignite outlier near Chambersburg, PA.
Jurassic	Extensive erosion of Appalachian Mountains; up to 9 km thick sediment wedge on the continental shelf and slope.

Table 3. (Continued)

Time and Climate	Features
Early Jurassic (Arid to semi-arid)	
Triassic (Arid to semi-arid)	Rifting; development of valleys.
Permian	Appalachian Orogeny; extensive folding and faulting of the Ridge and VAlley region.
Early Permian	Possible deposition of terrestrial sediments.
Pennsylvanian (Temperate)	Deposition of coal measures of marine, brackish and fresh water origin.
Cambrian-Pennsylvanian	Development of Appalachian geosyncline.



Stop localities for Field Trip #3, plotted on Landsat Image of part of Central Pennsylvania (scene E 1243-15253, band 7, of March 23, 1973).

FIELD TRIP #3

FIELD GUIDE - ON THE SOILS AND GEOMORPHIC EVOLUTION OF NITTANY VALLEY

R. R. Parizek and W. B. White
Field Leaders

0.0 Holiday Inn. Turn left and follow Atherton Street, Route 322 to the northwest.

0.6 Westerly Parkway. These shallow tributary valleys to the Spring Creek drainage are all underdrained even before the installation of streets and storm drains. The storm drain systems in this part of State College are discontinuous. They drain into sinkholes. Natural conduit systems are used as an essential part of the storm drain. The water ultimately resurges as springs in Walnut Spring Park and in Thompson Spring on the eastern side of State College.

1.5 College Avenue. Cross Route 26 and continue west on Route 322.

2.7 Big Hollow. The main underdrained tributary of the State College area. Water table maps indicate a groundwater trough paralleling Big Hollow some 100 feet below the valley bottom.

4.2 Park Forest Village. We are here crossing the northern end of Gatesburg Ridge, underlain by the Gatesburg dolomite with thick, low-organic-content residual soils. The site of the former State College land fill lies on these residual soils just to the right of the Highway. The stands of oak that shade the Park Forest Village housing development are typical of the plant cover supported by the Gatesburg soils.

5.3 Cross State College By-Pass.

7.5 Matternville. Cross Route 550 and continue up Bald Eagle Ridge on Route 322. The rocks exposed are first the Oswego or Bald Eagle sandstone and then the red Juniata shales and siltstones.

8.3 Skytop. STOP 1.

From this vantage point one sees the transition from the Valley and Ridge Province to the Allegheny Plateau Province. The bedrock under foot is the Silurian Tuscarora sandstone. The lower Devonian Helderberg limestones and Oriskany sandstone crop out near the base of the ridge. Bald Eagle Creek flows on a valley of Devonian shales. The foreridges of the Allegheny front are the Catskill red shale and the Pocono sandstone. The top of the escarpment is supported by the Pennsylvanian Pottsville sandstone.

The escarpment itself is a major geomorphic feature that can be traced, at continuously rising elevation southward through Maryland and West Virginia to become the Cumberland Escarpment of Tennessee.

From Skytop, return eastward on Route 322.

9.1 Intersection with route 550. Turn right onto Route 550.

11.6 Turn left onto dirt road that crosses the valley toward Gatesburg Ridge.

12.7 Clay pits and related mining activity in Gatesburg residual soils. [STOP 2].
Return along same route.

13.8 Route 550. Turn right.

16.3 Route 322 intersection. Turn right, eastward.

18.2 Leave Route 322 and continue straight ahead onto the State College By-Pass.

The By-Pass is cut through the Gatesburg Ridge. Residual soils are exposed where construction work is underway on a new interchange.

22.6 Exit the By-Pass onto Park Avenue Extension (Pennsylvania State University Exit).

The lands along Park Avenue are part of the University farms underlain mainly by Ordovician dolomites.

23.8 Traffic light. Turn right onto Fox Hollow Road.

24.6 Turn left onto blacktop road just past National Guard Armory and before crossing under the By-Pass.

25.2 Turn right onto gravel road just past the fire school.

25.3 Take left fork of gravel road.

25.4 Farm gate on right. Barrow pit exposing residual soils is on the hillside beyond [STOP 3].

The bulk of the soil exposed at this location is residual from the Gatesburg formation. Relict bedding structures are visable. Late Pleistocene soils are visible as a thin veneer capping the sequence.

Backtrack to Fox Hollow Road.

25.5 Turn left on blacktop road. The University water supply is obtained from wells drilled in the Gatesburg aquifer along this hollow and its tributaries. In general, the Gatesburg is a productive aquifer in the State College Region because of its high primary permeability through sandy units and vuggy openings.

26.1 Fox Hollow Road. Turn left.

27.1 Turn left into Toftrees Development. Follow the main street, Toftrees Avenue.

27.4 Turn right onto gravel lane which leads immediately to the gate at State Gamelands site [STOP 4].

This site is one of two used for disposing of sewage plant effluent by the Living Filter concept. The Gatesburg soils are thick, the water table is some 200 feet below the land surface. The shallow depressions are slump features in the soils which hold perched ponds of water.

Backtrack.

27.6 Fox Hollow Road. Turn right.

29.5 Traffic light at Park Avenue Extension. Turn left.

30.3 Turn right onto State College By-Pass.

View of Nittany Mountain is good from the By-Pass. The nose of Nittany Mountain is the steep outer slope formed by the northeast-plunging syncline.

31.6 Traffic light at intersection with Route 26, Benner Pike. Turn left.

32.3 Traffic light. Turn left onto Houserville Road.

32.8 Turn left into Spring Creek Park.

32.9 Spring Creek Park. LUNCH. [STOP 5].

Spring Creek here flows on a narrow flood plain not far below the valley uplands. A terrace level thought to be Sangamon age lies only a short distance above the Creek. A buried soil can be observed here.

Return to Benner Pike.

33.5 Traffic light at Benner Pike. Turn left.

34.4 Y-intersection at Nittany Mall shopping center. Keep left on Route 150.

The Nittany Mall provides an interesting example of environmental problems with land development in karst areas. The large extent of roof and parking lot concentrates the runoff and enhances sinkhole development at the margins of the area. Much of the runoff is concentrated in a sinkhole visable in the arms of the Y of the highway intersection. Other sinkholes parallel the flank of Nittany Mountain. Many of these are filled with trash. The sinkholes provide pathways for the injection of solid wastes into the conduit system carrying groundwater below.

Benner Pike crosses the valley uplands diagonally between the Nittany Mall and Bellefonte. Here the Gatesburg Ridge is not prominent and there is a uniform upland surface, mostly on the Ordovician dolomites spanning the valley between Nittany Mountain and Bald Eagle Ridge.

38.2 Turn left on black top road to Fisherman's Paradise.

38.8 T-intersection. Turn left.

39.4 Parking area at Fisherman's Paradise [STOP 6].

Spring Creek has here downcut into a deep gorge below the flood plain at Spring Creek Park. The creek provides a pronounced trough toward which groundwater discharges as a series of springs emerging from the carbonates along the banks of the creek. Benner Spring is one of the larger.

Return down Spring Creek.

40.0 Continue straight at the intersection.

The floor of the Spring Creek gorge has bits of floodplain here and there as the valley widens and narrows. This suggests a stable location of the creek for some time, presumably due to the damming of the creek by the resistant Tuscarora quartzite which the stream must cross at Milesburg Gap. The creek is spring-fed and thus maintains a stable flow throughout the year.

41.7 Intersection. Turn right.

42.1 Intersection with Route 550. Turn right.

42.9 Intersection with Route 150 in Bellefonte. Turn left.

43.3 Big Spring rise pool behind the wall on the left. Big Spring is the largest carbonate spring in the valley with an estimated discharge of 30 cfs. The water has unusually low hardness for a carbonate spring, 130 ppm as $CaCO_3$. The spring provides the water supply for Bellefonte and for the Corning Glass Works near the Nittany Mall.

43.4 Traffic light at High Street. Continue straight ahead.

Beyond the High Street intersection, Spring Creek flows in an artificial channel just to the left of the roadway. Comparison of the flow seen here with the flow in Spring Creek gorge gives some impression of the contribution of Big Spring and Logan Branch to the overall runoff from the valley.

43.8 Intersection. Continue straight ahead.

44.9 Entering Milesburg Gap through Bald Eagle Ridge. There are only six feet or so of alluvial sediment overlying bedrock in the stream channel. Steeply dipping beds of Bald Eagle sandstone and Tuscarora quartzite plunge below stream level and form a dam which resists downcutting of the stream behind it.

45.3 Fill bank exposing colluvial soils. Pull off on right [STOP 7].

Return through Milesburg Gap.

46.7 Intersection with Rt. 144. Keep right on Route 150.

47.2 High Street intersection. Straight ahead.

47.4 Intersection. Turn left to Route 144.

47.5 Intersection. Continue straight ahead on Route 144.

48.3 Cerro metal plant. The cut to the left of the road is one of the best exposures of the Gatesburg dolomite. The characteristic sandy beds and vuggy texture are clearly visable here.

49.3 Axemann Spring to left of road in stone springhouse. This has been a water supply spring in the past and may be used again.

Route 144 follows the valley of Logan Branch which here appears to have a meandering course. Close examination of the stream pattern on topographic maps or air photographs reveals that the stream is flowing diagonal to the regional fracture system and thus pursues a zig-zag course with individual straight valley segments oriented along fracture traces.

Like the other tributaries to Spring Creek, Logan Branch is spring-fed. The source of the Branch is a group of springs west of Pleasant Gap.

51.7 Traffic light at intersection with Route 26 in Pleasant Gap. Turn left on Route 26.

52.2 Traffic light. Turn right on Harrison Road.

52.5 Hildas Beauty Parlor pull-off on left [STOP 8].

Gap Run sinks in its bed behind the Beauty Parlor. Like most other swallow holes in the valley, Gap Run sinks into the Champlainian limestone. There is no stream channel downstream from the swallow hole. This implies that all sediment transported from the mountain flanks must be carried by the conduit system. It also has implications for the behavior of large floods that overwhelm the swallow hole.

Continue up Harrison Road toward Nittany Mountain.

52.8 Intersection with Route 144. Turn left up the gap.

In spite of the unreliability of the supply, mountain streams continue to be prized as water supplies because of low hardness and good chemical quality. The pipes seen along the highway are part of the collector system for the Pleasant Gap Water Company. The actual collectors are in the small tributary valleys away from the influence of the highway.

Much of this roadway was washed out in the Hurricane Agnes storm of 1972. The concrete retaining walls and new fill structures were emplaced at that time.

55.1 Nittany Mountain summit. There is an excellent view into Penns Valley in an interfluve area between the Spring Creek drainage to the west and the Penns Creek drainage to the east. The valley floor is beveled across folded carbonate rocks and represents the Harrisburg surface at about 1300 feet elevation. The excellent preservation of the surface is due partly to its location in an interfluve area and partly to karstic drainage so that there is no dissection of the valley uplands.

56.3 Borough of Centre Hall. There is a typical large sinkhole in the Champlainian limestone to the left of the highway at the foot of the mountain.

57.7 Traffic light. Intersection with Route 45 in Old Fort. Turn right onto Route 45.

Somewhere along the highway is the subsurface divide. To the east, underground flow is to the Penns Cave Rising or to the Spring Mills spring some five miles away. To the west the drainage rises in the Linden Hall springs. A few miles west of Old Fort the trace of a westward-trending dry valley appears.

62.6 Upper Linden Hall Springs to the right. These have shallow rise pools in the valley floor with no enterable conduits. The discharge from the springs is part of the Spring Creek drainage.

65.1 Intersection with Route 322 in Boalsburg. Turn right and follow Route 322 back to State College.

68.0 Holiday Inn on left. End of field trip.

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