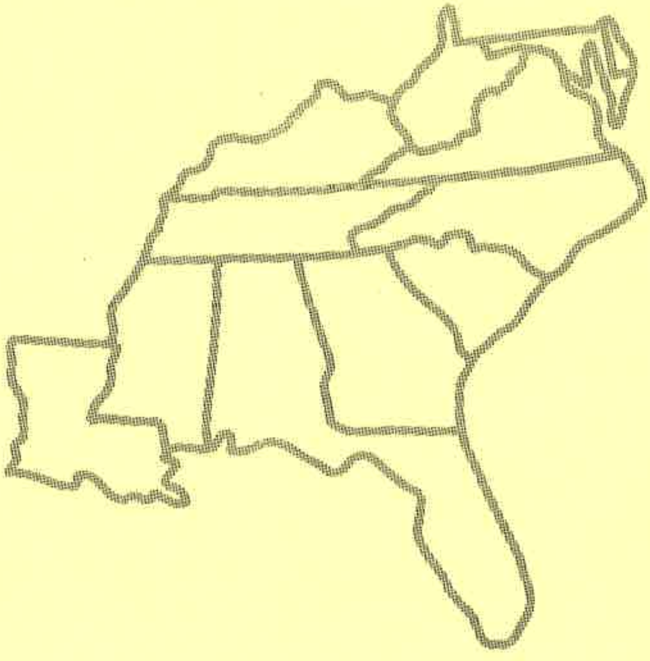
SOUTHEASTERN GEOLOGY



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THREE-STAGE MODEL OF BRITTLE DEFORMATION IN THE CENTRAL APPALACHIANS

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ABSTRACT

Recent fieldwork in the central Appalachians shows the applicability of a three-stage model of brittle deformation. Distinct minor structures form at specific times during overall horizontal shortening and map-scale folding. Stage I contraction faults shorten horizontal or gently dipping beds and are closely associated with wedging. Stage II uplimb thrust faults shorten folded beds hingeward. Stage III extension faults and extension fractures lengthen steeply dipping to overturned beds. Opening or shearing directions of fractures and crosscutting relationships of faults, other fractures, and stylolites allow distinction of the three stages. Stage III may have produced fracture porosity and permeability on steep limbs of anticlines. We propose a three-stage model of outcrop-scale fault and other fracture development for steeply dipping to overturned beds in eastern West Virginia and adjacent areas. Less rotated beds may show at least two of the stages. Kinematic analysis of faults and other fractures allows differentiation of the three stages.

Acknowledgments

We wish to thank Wallace de Witt, Jr. and Clinton D. A. Dahlstrom, among many others, with whom we have had useful discussions in the field. Comments of de Witt, M. T. Heald, and R. C. Shumaker improved the manuscript. Many of the concepts herein are influenced by the work of R. A. Price, who cannot be held accountable for our conclusions with respect to the Appalachians. Early versions of the manuscript were prepared with a modification of WYLBUR, a text-editing system developed at the Stanford University Computation Center.

STRUCTURAL STYLE

The central Appalachian foreland has many detached anticlines but few outcropping major thrust faults. The anticlines are interpreted as active features generated mainly by duplication of strata by ramping of underlying thrust faults, by splay faults, and by ductile flow of shale-rich intervals into anticlinal crests (Perry and de Witt, 1977, Perry, 1978, Wheeler, 1975). The synclines are regarded as passive features resulting from anticlinal growth in adjacent rocks, rather than from active downbuckling (Gwinn, 1964).

We deal with the sequence of outcrop-scale structures that formed at specific times during overall southeast-northwest horizontal shortening and vertical extension. We assume that each individual structure formed in an orientation that allowed it to accommodate some of that horizontal shortening, or vertical extension, or both.

The first-order anticlines (Nickelsen, 1963, p. 16) of the western Valley and Ridge and the Allegheny Plateau provinces formed predominantly by flexural slip folding (Faill, 1969), rather than by passive or flexural flow folding (Gair, 1950). Interlayering of shales and evaporites with sandstones, siltstones or limestones allows slip between units or low relative ductility. Stage I structures form before folding, or early during folding, when the angle between bedding and the maximum principal compressive stress is less than about 10 degrees. Experiments at room temperature and confining pressures to 2000 bars (equivalent to about 6 km in depth) suggest that slip parallel to bedding is possible when the angle between the maximum principal compressive stress and bedding ranges from 10 to 60 degrees (Price, 1967). Stage II structures form in this range of limb dips (see below). Stage III structures form in response to additional horizontal shortening, late in or after folding, when steep limb dips preclude further slip parallel to bedding. These structures form when the angle between bedding and the maximum principal compressive stress exceeds about 60 degrees. These conclusions are reinforced by the finite-element derived stress-history of folding (Dieterich and Carter, 1969).

Price (1967) described extension and contraction structures from the Canadian Rocky Mountains using the fault terminology of Norris (1958). We report the same types of structures from the central Appalachians, placing them in a relative time sequence and describing simple field criteria for differentiating among stages.

Stage I Structures

Cloos (1964) first recognized and named the prefolding wedges common in the central Appalachians. Wedges are the wedge-shaped ends of small fault blocks in which the bounding contraction faults (Norris, 1958) form angles of 30 degrees or less to bedding. Examples provided by Cloos (1964, figs. 2, 3, 4, and 6) involve brittle layers (sandstone and limestone) which have been "sheared, wedged, and telescoped together" in a more ductile medium (shale). This process of contraction faulting shortened the stratigraphic section in a northwest-southeast direction and thickened it perpendicular to bedding prior to or early during folding. Such contraction faults can form dipping in either direction (northwest or southeast).

Concerning the Canadian Rockies, Price (1967) writes that if layering was planar and inclined at a low angle to the maximum principal compressive stress, that stress' trajectories would tend to parallel layering, and subsequent failure would take the form of contraction faults acting to shorten layers. The resulting geometry of brittle beds in a more ductile matrix is that of Cloos' wedges. Compound wedges involving a series of brittle beds which have been telescoped together are shown by Cloos (1964, fig. 7) and Perry and de Witt (1977, fig. 10). Price (1967) used the term contraction faults to include all faults that produce a shortening in the plane of the bedding, thus including uplimb thrust faults (see below).

In wedged beds later rotated to vertical by folding, the bounding contraction faults record normal-fault separation at a low angle to bedding. In cratonward-facing folds (asymmetric to the northwest in the Appalachians), such stage I faults normally show a downlimb sense of displacement (Figure 1). They did not form as normal faults in their present orientation, because (1) the necessary northwest-southeast horizontal extension is inconsistent with central Appalachian structural style, and (2) their formation in a compressional regime is indicated by drag features and absence of associated extension features.

Prefolding fractures have been recognized in the central Appalachians (Dean and Kulander, 1977). These fractures may predate the contraction faults, because such fractures are offset by these faults and by bed-parallel compaction stylolites that are inferred to have formed under overburden stress equal to maximum principal compressive stress, when bedding was horizontal.

Prefolding calcite-filled fractures can be recognized if orientations of calcite fibers record shearing on the fractures. If the fractures are rotated to their original (prefolding) orientations, the net growth direction of the fibers should be about 30 degrees from the (horizontal) maximum principal compressive stress.

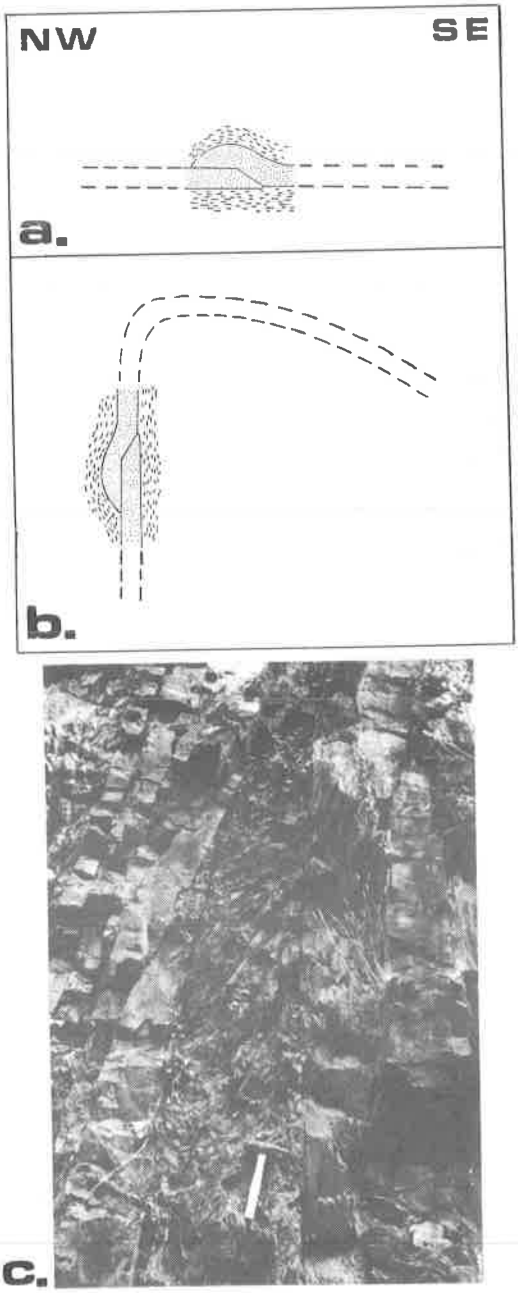

Figure 1. Contraction fault (stage I) formed wedges prior to folding: (a) prefolding attitude, (b) postfolding attitude on northwest limb of west-facing anticline, (c) present attitude on northwest limb of major west-facing fold near Woodmont, Maryland. The angle between the bedding and fault planes and the size of the associated fold is variable.

Stage II Structures

Stage II structures formed during folding. They record relative reverse movement toward the anticlinal hinge as part of the Appalachian fold's internal adjustments to southeast-northwest horizontal shortening, and to fold growth by flexural mechanisms. Perry and de Witt (1977) describe and define uplimb thrust faults, the hanging walls of which move up the limb of the anticline and away from the axis of the adjacent syncline (Figure 2). Similarly, Gair's (1950) out-of-syncline thrusts, and Gwinn's (1964) symmetrical thrust faults are northwest- and southeast-dipping reverse faults that resolve space problems in cores of anticlines. In concentric folding, upward and inward motion on anticlines' flanks, and flexural slip above ductile rocks, thrust strut-like brittle beds of the limbs over passive anticlinal crests where flexural slip is inhibited. Perry (1971) mapped southeast-dipping apparent normal faults at a low angle to bedding in vertical beds on the northwest limb of the Wills Mountain anticline in Pendleton County, West Virginia. He interprets these as originally northwest-dipping uplimb thrust faults (Perry, 1971, 1978), later rotated by continued growth and asymmetric development of the anticline. Rotated uplimb thrust faults are also present on the nearly vertical northwest limb of the Cacapon Mountain anticline in Maryland (Perry and de Witt, 1977, p. 32).

Stage II structures shorten beds and facilitate anticlinal growth. Both can occur contemporaneously in a planar, mechanically anisotropic medium if net transport is mostly toward the anticlinal hinge. Some rotation of stage II structures occurs if they form early, at low limb dips (Root, 1973). Uplimb thrusts also show a wedge-like geometry and may be difficult to distinguish from stage I wedges. Uplimb thrusts tend to cut many beds and die out in shale flowage (Gair, 1950), folds (Gwinn, 1964), or bed parallel slip (Price, 1964). Wedges tend to be small because they represent the adjustment to shortening of a single or a few strut-like beds. Wedges can show a downlimb sense of displacement on northwest limbs of anticlines. Uplimb thrusts tend to cut more beds because they represent adjustment to shortening of the entire fold. Uplimb thrusts always show hingeward displacement of the upper (or outer) fault block.

As uplimb thrust faults on limbs steepen during growth of the anticline, the normal stress across the thrust surface and the resulting friction increase until the fault locks (Gwinn, 1964). Further tightening of the anticline can produce another fault or faults on the limbs. Continued growth of the anticline, beyond that which flexural slip is capable of relieving, causes onset of stage III.

Stage III Structures

Stage III structures begin to form when beds rotate so far towards

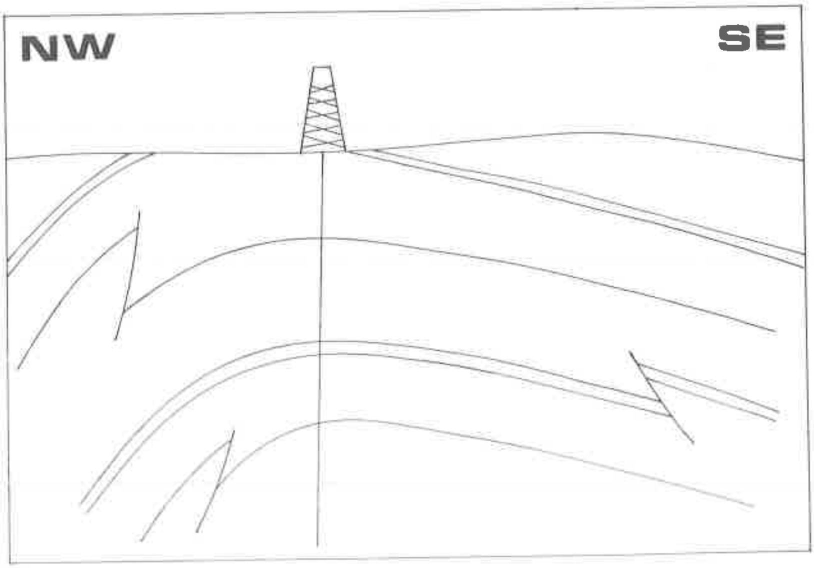

Figure 2. Southeast- and northwest-dipping uplimb thrust faults (stage II).

the vertical that the effect of vertical extension exceeds that of horizontal shortening. Further growth of the anticline can then only occur by bed-parallel extension. Stages I and II both involve bed-parallel contraction, and so can overlap in time. Because the change from bed-parallel contraction to bed-parallel extension is a discrete event, stages II and III are unlikely to overlap, and their structures should be readily distinguishable. Stage III structures are most easily recognized on and are especially characteristic of steeply dipping to overturned beds.

Norris (1958, 1964) defined extension faults as those that result in elongation in the plane of the layering. He reported extension faulting in otherwise ductile beds (carbonaceous shales) dipping 10 to 25 degrees. Price (1964, 1967), Perry (1971, 1978) and this paper describe extension faults in brittle, steeply dipping beds. Norris and Price noted that if the layering rotated externally in the course of flexural-slip folding until it was at a high angle to the maximum principal compressive stress, then the trajectories of that stress would tend to become perpendicular to the layering and subsequent brittle failure would take the form of extension faults. This rotation of the stress trajectories is shown by the finite-element modelling of Dieterich and Carter (1969). Price (1967) has found that extension faults tend to intersect bedding at about 70 degrees.

Extension faults show net bed-parallel lenthening unique in the kinematic history of a fold, thus allowing easy recognition. Characteristically, extension faults are low-angle, northwest- or southeast-dipping

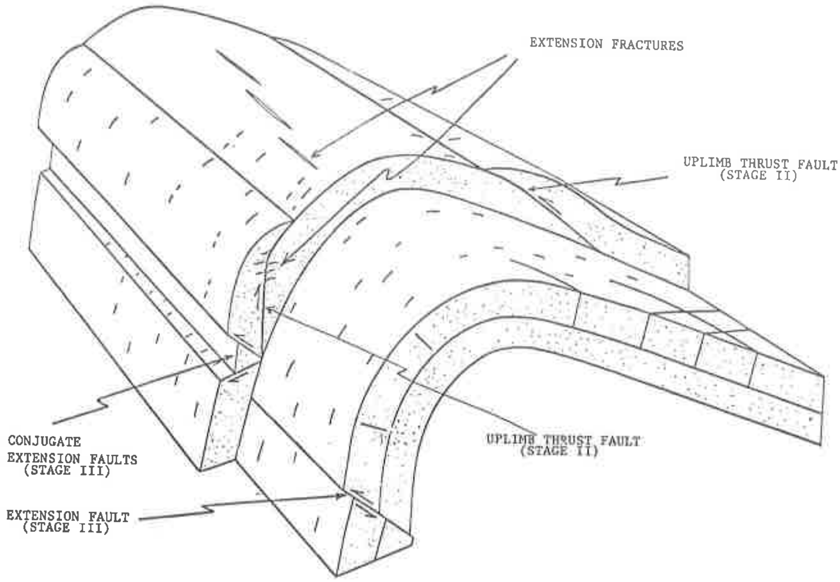

Figure 3. Relationships of stage II uplimb thrust faults to stage III extension faults on an asymmetric fold (modified from Perry and de Witt, 1977, after Price 1967). Extension fracturing may be associated with both stages II and III.

reverse faults on steep or overturned limbs (Figure 3). Extension fractures which show bed-parallel lengthening and form normal to bedding are stage III structures. Extension faults have not undergone a significant amount of later rotation. If they formed prior to folding, they would have formed as high angle normal faults, which are inconsistent with Appalachian compressive deformation. Specifically, extension faults cut contraction faults of stages I and II (Perry, 1971, Perry and de Witt, 1977) and are therefore later features. In extreme cases, map-scale extension faults can lead to overthrust west limbs of anticlines (Dennison, writ. comm., 1978).

With the possible exception of some systematic joints, stage III structures are the latest recognizeable effects of brittle, fold-related deformation in the central Appalachians. Because they are least likely to be filled by vein material or closed during later deformation, they are a possible target for gas exploration in fractured rock.

FIELD APPLICATIONS

Recent fieldwork has shown the applicability of our three-stage model of brittle deformation in steeply dipping beds in the central

Appalachians. Nine exposures of Ordovician through Mississippian limestones, sandstones, siltstones, and shales of West Virginia, Maryland and Virginia have been interpreted in terms of the model to yield a sequence of events consistent with Appalachian structural style. Slip-senses of minor faults are usually apparent and show either stage I contraction faulting, stage II crestward slip, or stage III bed parallel extension, depending on the angle between beds and horizontal shortening when failure occurred. Directions of fracture openings are ambiguous unless the fractures are filled with fibrous calcite, cut across compaction or tectonic stylolites, or are slickensided. Durney and Ramsay (1973) showed that crystal fibers in calcite fracture-fillings track the direction of opening of the fracture. If the fracture opened by simple dilatancy, then successive orientations of incremental extension during fiber growth can be determined. If the direction of fracture opening is parallel or at a small angle to bedding, then the fracture is unequivocally a stage III feature. Fractures whose opening directions have components of crestward shearing can be either stage I or stage II.

An excellent exposure showing examples of all three stages of brittle deformation is the cut in Silurian limestones along the Baltimore and Ohio Railroad, on the northwest flank of the Wills Mountain anticline at Pinto, Allegany County, Maryland.

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1200 M. Y. -OLD GNEISSES IN THE BLUE RIDGE PROVINCE OF NORTH AND SOUTH CAROLINA

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ABSTRACT

Rb-Sr whole-rock analyses indicate the existence of approximately 1200 m. y. -old basement rocks in the Blue Ridge Province of central and southern North Carolina, plus adjacent parts of South Carolina. Migmatitic granitic gneiss from the Mars Hill quadrangle, North Carolina, has an age of 1183 $\pm$ 65 m. y. with an initial Sr^{87}/Sr^{86} ratio of 0.7069 $\pm$ 0.0009. Toxaway Gneiss, collected near the North and South Carolina state boundary, is 1203 $\pm$ 54 m. y. -old and has an initial Sr^{87}/Sr^{86} ratio of 0.7016 $\pm$ 0.0021. Some of the Toxaway samples may have undergone partial isotopic re-equilibration during a mid-Paleozoic metamorphic event.

INTRODUCTION

Some of the granitic gneisses of the Blue Ridge province of the southern Appalachians are known to be 1300 to 1000 m. y. -old (Tilton and others, 1960; Davis and others, 1962; Dietrich and others, 1969; Fullagar and Odom, 1973). However, the areal extent of rocks of this age is still poorly known. This study reports new Rb-Sr age results for

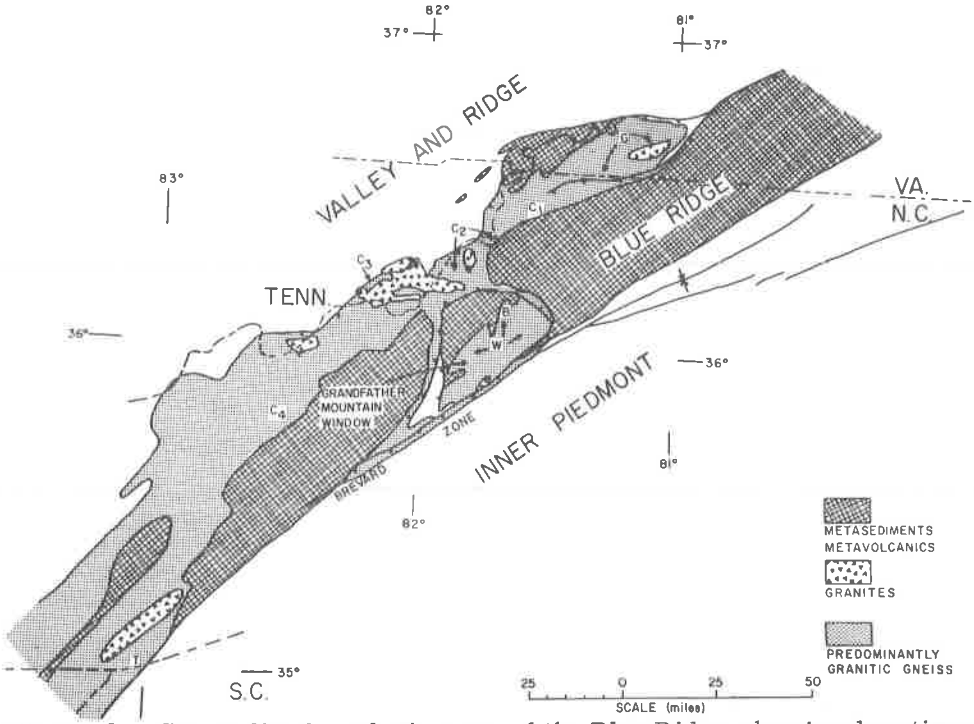

Figure 1. Generalized geologic map of the Blue Ridge showing locations of 1200 to 1000 m. y. -old gneisses: B, Blowing Rock Gneiss, N. C. ; C_1, Cranberry Gneiss, Ashe Co. , N. C. ; C_2, Cranberry Gneiss, Watauga Co. , N. C. , and Johnson Co. , Tenn. ; C_3, Cranberry Gneiss, Pardee Point, Tenn. ; G, Grayson Gneiss, Va. ; C_4, Migmatite, Mars Hill quadrangle, N. C. ; T, Toxaway Gneiss, North and South Carolina; W, Wilson Creek Gneiss, N. C. See Table 3 for ages.

migmatitic granitic gneisses from the Mars Hill quadrangle in North Carolina and the Toxaway Gneiss in southern North Carolina and adjacent South Carolina. Figure 1 shows these locations plus the locations of other gneisses which have ages of 1300 to 1000 m. y.

Acknowledgments

This study was supported in part by funds from the North Carolina Department of Natural and Economic Resources, Geology and Mineral Resources Section, the South Carolina Division of Geology and TVA. The manuscript was reviewed by A. L. Odom and L. S. Wiener.

Table 1. Rb-Sr Analytical Data for Migmatite from Mars Hill Quadrangle.

Sample No.	$(Sr^{87}/Sr^{86})_N$	Rb ppm	Sr ppm	Rb^{87}/Sr^{86}
1430	0.7371	96.4	153.9	1.819
1431	0.7144	69.1	414.8	0.483
1432	0.7218	109.1	378.5	0.836
1433	0.7266	150.8	404.9	1.080
1434	0.7186	114.8	425.4	0.782
1435	0.7134	82.4	649.9	0.367
1436	0.7170	102.9	503.2	0.593
1437	0.7047	25.6	749.7	0.099
1438	0.7049	21.5	764.9	0.081

Table 2. Rb-Sr Analytical Data for the Toxaway Gneiss

Sample No.	$(Sr^{87}/Sr^{86})_N$	Rb ppm	Sr ppm	Rb^{87}/Sr^{86}
Whitewater Falls, N. C.				
T-56	0.7464	186.5	201.6	2.689
1608	0.7364	180.2	259.2	2.018
1610	0.7489	258.7	235.2	3.197
1612	0.7419	189.7	235.5	2.340
Thompson River, S. C.				
T-66, A	0.7304	138.3	246.6	1.626
T-42, B	0.7480	153.4	176.0	2.534
S. C. Highway 413 E				
1401	0.7420	200.3	216.3	2.690
1402	0.7463	247.7	215.1	3.346
1602	0.7434	210.5	240.5	2.543
1603	0.7611	240.0	196.4	3.555
1604	0.7829	216.2	136.0	4.635

ANALYTICAL PROCEDURES

Standard techniques were used for sample preparation and analyses for Sr and Rb concentrations and Sr isotopic compositions. More specific information on procedures is given in Fullagar and Odom (1973). Analytical data are given in Tables 1 and 2. The $(Sr^{87}/Sr^{86})_N$ ratios are reported relative to a value of 0.7080 for the Eimer and Amend $SrCO_3$ reference sample. Analytical uncertainties (one-standard-deviation) are estimated to be less than 1.0 percent for Rb^{87}/Sr^{86} ratios and less than 0.05 percent for $(Sr^{87}/Sr^{86})_N$ measurements. The size of the data points in Figures 2 and 3 represents uncertainties of

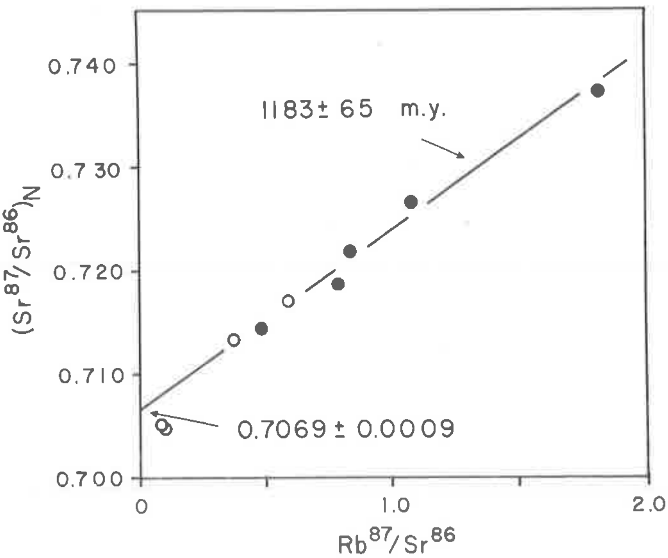

Figure 2. Rb-Sr isochron diagram for samples from migmatite (granitic gneiss), Mars Hill quadrangle, N. C. Solid symbols represent samples from quarry no. 17; open symbols represent samples from quarry no. 21. See text for discussion.

approximately two-standard-deviations. Ages and initial Sr^{87}/Sr^{86} ratios were calculated with the least-squares cubic treatment of data suggested by York (1966). A Rb^{87} decay constant of $1.419 \times 10^{-11} yr^{-1}$ (Davis and others, 1977) was used to calculate the new ages reported; published ages (Table 3) were recalculated using this decay constant. All age and initial Sr^{87}/Sr^{86} uncertainties quoted in this paper represent one-standard-deviation.

MARS HILL QUADRANGLE, NORTH CAROLINA

This quadrangle is located in the Blue Ridge of western North Carolina, partly in southeastern Madison County and partly in northern Buncombe County. Merschat (1977) has reported on the geology and the mineral resources of this quadrangle. Two of Merschat's earlier Precambrian map units originally included within Keith's (1904) Cranberry Granite and Roan Gneiss were sampled and analyzed. Seven samples of biotite granitic gneiss (map unit bgg_1) yield a scatter of data points when plotted on a Rb-Sr isochron diagram. As the scatter is too great to permit a meaningful interpretation, these data are not given in this paper and this unit will not be discussed further. Nine samples of biotite-hornblende migmatite (map unit bhm_2) were analyzed and the data are given in Table 1.

The migmatitic unit consists mainly of interlayered biotite-hornblende gneiss and granitic gneiss. For this study, granitic gneiss layers were sampled in two quarries. Samples 1430 to 1434 came from quarry no. 17 (Merschat, 1977; North Carolina coordinates 762,050 N., 948,350 E.). A single thin-section from quarry no. 17 displays both granoblastic and lepidoblastic textures. This thin-section is composed of 52 percent K-feldspar (orthoclase and microcline), 17 percent plagioclase, 30 percent quartz and 1 percent biotite, and is considered typical of the analyzed samples. Samples 1435 to 1438 are from quarry no. 21 (Merschat, 1977; North Carolina coordinates 762,250 N., 953,900 E.). These samples were somewhat more mafic than those from quarry no. 17. The granitic gneiss in this quarry is in close proximity to Bakersville Metagabbro, which crops out high on the quarry wall.

Data for the granitic gneiss from the Mars Hill quadrangle are plotted in Figure 2. Using York's (1966) regression calculation all nine data points indicate an age of 1320 $\pm$ 76 m. y., with an initial Sr^{87}/Sr^{86} ratio of 0.7047 $\pm$ 0.0009. Though this could be the correct age for these rocks, we favor an alternate interpretation of the data. The samples collected from quarry no. 21 (1435-1438) were located no more than several tens of feet from the contact with the Bakersville Metagabbro. Thus, it is possible that the granitic gneiss might have been altered by the intrusive Bakersville Metagabbro. The high Sr content of samples 1435-1438, especially 1437 and 1438 (~750 ppm) suggests the possibility that these granitic gneiss samples were altered by the gabbro. Accordingly, data for samples 1437 and 1438 were excluded from the regression analysis. The age which resulted was 1183 $\pm$ 65 m. y., with an initial Sr^{87}/Sr^{86} ratio of 0.7069 $\pm$ 0.0009; these are the values given in Figure 2. It is obvious from Figure 2 that these data points are not perfectly co-linear, as the data points deviate somewhat from the isochron. This scatter may be the consequence of samples having different initial Sr^{87}/Sr^{86} ratios, or, Rb-Sr systems of the whole-rock samples may have been altered, perhaps during a Paleozoic metamorphic event(s). It could be argued that all of the samples from quarry no. 21 were altered by intrusion of the gabbro. If we consider just the samples from quarry no. 17, the age obtained is 1201 $\pm$ 96 m. y. and the initial Sr^{87}/Sr^{86} ratio is 0.7065 $\pm$ 0.0015. This age is not significantly different than the one given above and on Figure 2. Regardless of how the data are interpreted the age for the granitic gneiss is approximately 1300 to 1200 m. y. The initial Sr^{87}/Sr^{86} ratio of the granitic gneiss is sufficiently high (0.705-0.707) to indicate that this rock probably was produced from an older sialic crust, or at least was contaminated by an older sialic crust (see Fullagar and Odom, 1973, for discussion).

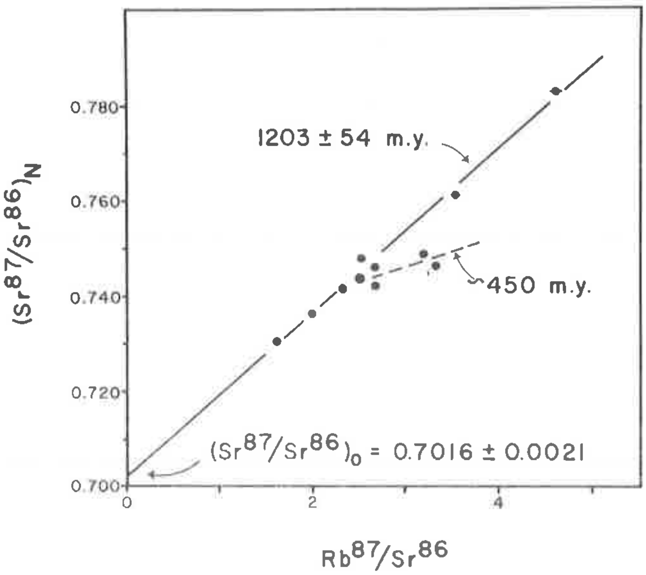

Figure 3. Rb-Sr isochron diagram for samples of Toxaway Gneiss, North and South Carolina. See text for discussion.

TOXAWAY GNEISS, NORTH AND SOUTH CAROLINA

The Toxaway Gneiss is exposed in the core of the Toxaway dome which is located in the Blue Ridge province of North and South Carolina, just northwest of the Brevard zone (Hatcher, 1977). The samples of Toxaway Gneiss that were analyzed came from two locations in the Cashiers quadrangle (outcrops on S. C. highway 413 E; Whitewater Falls, N. C.) and one location in the Reid quadrangle (near Thompson River, S. C.).

As described by Hatcher (1977), the Toxaway Gneiss is typically a granitic to quartz monzonitic gneiss with alternating quartz-feldspar and biotite-rich bands. Microcline generally is the dominant feldspar, and biotite is the dominant mafic constituent. Hatcher suggests this gneiss may be metasedimentary in origin.

Eleven samples of Toxaway Gneiss were analyzed and the results are plotted in Figure 3. Eight of the eleven samples indicate an age of 1203 $\pm$ 54 m. y. with an initial Sr^{87}/Sr^{86} ratio of 0. 7016 $\pm$ 0. 0021 As several of these eight samples plot slightly off the isochron, they may have had different initial Sr^{87}/Sr^{86} ratios or they may have been chemically (or isotopically) altered during a metamorphic episode. Nonetheless, because of the reasonably good fit of these eight data points to the isochron, we suggest that the age of this unit is approximately 1203 m. y. The low initial Sr^{87}/Sr^{86} ratio (0. 7016 $\pm$ 0. 0021) probably places some constraints on the origin of the Toxaway Gneiss. The Toxaway could be an igneous rock which was derived from the upper mantle (or lower crust), or it could be a metasedimentary rock which was produced from materials which had been part of the upper (sialic) crust for a relatively short period of time (perhaps less than 100 to 200 m. y.).

The Blue Ridge rocks in the area of the Toxaway dome have been subjected to at least one episode of progressive Paleozoic metamorphism (Hatcher, 1976; 1977). The peak of regional metamorphism, which raised these rocks to middle or upper portions of the amphibolite facies, probably occurred between 480 to 440 m. y. ago (Butler, 1972; Kish, 1975; Dallmeyer, 1975). For reference purposes, a 450 m. y. isochron has been drawn through the three data points in Figure 3 which plot significantly to the right of the 1203 m. y. isochron. Two of these three points (1401 and 1402, see Table 2) are from the same outcrop and would plot on a line (not shown) with a slope corresponding to an age of about 450 m. y. (The other three samples from this outcrop (1602, 1603 and 1604) plot on or nearly on the 1203 m.y. isochron shown in Figure 3.) The third sample that plots significantly to the right of the 1203 m. y. isochron is sample 1610; four samples were collected from this outcrop: T-56, 1608, 1610 and 1612. All but 1608 also would plot on or very close to a 450 m. y. isochron (not shown). Thus, some of the data are compatible with an interpretation that some of the rocks underwent re-equilibration of Sr isotopes during metamorphism approximately 450 m. y. ago. If this is what happened, re-equilibration must have been on a scale larger than the 5 to 10kg samples that were analyzed. It must be emphasized that the data do not provide independent evidence for a metamorphic event 450 m. y. ago. The data points exhibit too much scatter to permit us to use them as evidence of metamorphic re-equilibration of the Sr isotopic composition. Additional analyses, especially of pegmatites which probably formed near the thermal peak of metamorphism (Hatcher, 1976), might demonstrate that the rocks were affected by a metamorphic episode approximately 450 m. y. ago.

>1,000 M. Y. -OLD BASEMENT ROCKS IN THE BLUE RIDGE OF SOUTHERN APPALACHIANS

Results presented in this paper allow us to add two locations to the list of 1200 to 1000 m. y. -old basement rock occurrences in the southern Appalachian Blue Ridge (Table 3, Figure 1). (Data initially reported by Tilton and others (1960) and Dietrich and others (1969) are re-evaluated in Davis and others (1962) and Fullagar and Odom (1973), respectively, and are listed in Table 3 under the latter references). In addition to the ages listed in Table 3, Davis and others (1962) also analyzed zircons from Pardee Point, Tennessee, and Deyton Bend, North

Table 3. Summary of 1200 to 1000 M. Y.-Old Ages from the Blue Ridge of Southwestern Virginia, Tennessee, North Carolina and South Carolina.

Unit	Age m. y. /Method	$(Sr^{87}/Sr^{86})_o$	Reference
Blowing Rock Gneiss, N. C. (B)*	1050/U-Pb	---	Davis & others (1962)
Blowing Rock Gneiss, N. C. (B)	1006 ± 35/Rb-Sr	0.7077 ± 0.0007	Fullagar & Odom (1973)
Cranberry Gneiss, N. C. (C_1)	1227 ± 44/Rb-Sr	0.7043 ± 0.0007	Fullagar & Odom (1973)
Cranberry Gneiss, N. C.-Tenn. (C_2)	1042 ± 40/Rb-Sr	0.7075 ± 0.0009	Fullagar & Odom (1973)
Cranberry Gneiss, Tenn. (C_3)	1050/U-Pb	---	Davis & others (1962)
Grayson Gneiss, Va. (G)	1150 ± 14/Rb-Sr	0.7044 ± 0.0004	Fullagar & Odom (1973)
Migmatite, Mars Hill Quad, N. C. (C_4)	1183 ± 65/Rb-Sr	0.7069 ± 0.0009	This paper
Toxaway Gneiss, N. C.-S. C. (T)	1203 ± 54/Rb-Sr	0.7016 ± 0.0021	This paper
Wilson Creek Gneiss, N. C. (W)	1050/U-Pb	---	Davis & others (1962)

*Symbols are same as on Figure 1.

Carolina. The age they obtained was approximately 1300 m. y., but since the zircons were detrital, crystalline rocks need not have existed at those locations at that time.

Much additional geochronological work needs to be done in the Blue Ridge. We still know relatively little about the distribution of the older basement rocks. Though Table 3 suggests that these basement rocks are approximately 1200 or 1050 m. y.-old, many more analyses must be done to determine whether or not this apparent pattern is real.

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HOLOCENE GEOMORPHIC EVOLUTION OF A BARRIER-SALT MARSH SYSTEM, SW DELAWARE BAY

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ABSTRACT

Delaware Bay is one of the larger estuaries along the eastern coast of North America. The paleogeomorphology of a portion of the southwest coast of Delaware Bay at Cedar Creek-Slaughter Beach has been delineated from airphoto studies, surface sedimentological and geomorphic studies of present day coastal environments, and sedimentological and paleontological studies of subsurface sediments of the latter half of the Holocene Epoch. Radiocarbon data from sedimentary environmental lithosomes clearly indicate that estuarine conditions existed in the study area around 5000 B. P. Studies of the sediments and the shapes of the sedimentary bodies encountered in a drilling program in a small area show that it is possible to make paleogeographic reconstructions of coastal zone environments during the Holocene in the Cedar Creek-Slaughter Beach area. Critical elements in such a study include the identification of the pre-Holocene surface and its morphological variants, the identification of Holocene environmental lithosomes in terms of their three-dimensional configurations and sediment type, and the radiocarbon dating of basal salt marsh organic rich sediments and other salt marsh sediments to provide a time frame for the sedimentary environments. It is possible to predict a rate of coastal erosion of approximately one meter per year in the area, based on observation of ongoing coastal processes and the geologic record of the last half of the Holocene.

INTRODUCTION

Atlantic coastal Delaware lies on the northwestern flank of the Baltimore Canyon Trough geosyncline. Geosynclinal deposits occur from the Fall Zone in the northwest to the outer edge of the Atlantic continental shelf. In the axis of the geosyncline, sediments are thicker than 12,000 m. In the area of this study (Figure 1), the southwestern portion of a major estuary, Delaware Bay, geosynclinal sediments are approximately 2000 m thick. These sediments have been deposited in varied sedimentary environments since Jurassic time. Presently, the geosyncline is partially emerged as the Atlantic Coastal Plain and partially submerged as the Atlantic continental shelf. The entire emerged and submerged Coastal Plain is covered by Quaternary sediments usually less than 30 m thick. The majority of the sediments on the emerged coastal plain are probably Sangamon Age fluvial, estuarine-lagoonal and shallow marine sands and muds. On the continental shelf and in adjacent lagoons and estuaries, Holocene sediments occur in thicknesses of up to 30 m. During the latest Wisconsin Ice Age, sea-level dropped to more than 100 m below present. During the peak Wisconsin regression, a deeply incised valley system developed on the northwestern flank of the Baltimore Canyon Trough geosyncline. Details of the ancestral Delaware River and its tributary systems have been traced onto the outer Delaware shelf (Swift, 1974 and Twitchell and others, 1977). Studies by Emery (1967), Emery and Garrison (1967) and Whitmore and others (1967) have identified coastal-marine sediments of Holocene age approximately 100 km offshore on the outer continental shelf. Their evidences included marsh deposits, the presence of *Crassostrea virginica*- the common oyster, and abundant remains of mastodons and mammoths.

Sea level rose relative to land at a fairly rapid rate in the early and middle portions of the Holocene Epoch and at ever decreasing rates toward the end of the Holocene to present time (Figure 2). Studies of relative sea level rise, utilizing over 80 radiocarbon dates, were used to develop a local relative sea level rise curve for the Delaware coasts (Kraft, 1976a, Kraft, 1976b, and Belknap and Kraft, 1977). In addition, Hicks and Crosby (1974), based on tidal gauge readings, show that the rate of relative sea level rise of the past half century has dramatically increased in the study area. They noted that in an eight year period in the 1960's sea level rose approximately 8.4 cm. It is not yet known what this observation means. It may be a fluctuation on a longer term average or it may be a change in cause and effect with a significantly higher rate of sea level rise to be projected into the future. Studies by Elliott (1972), Kraft and others (1973), Kraft and John (1976), and Strom (1972) provide supporting evidence for the interpretations made herein. However, they do not answer the question raised by Hicks and Crosby regarding a possible dramatic acceleration in rate of sea level rise in present times.

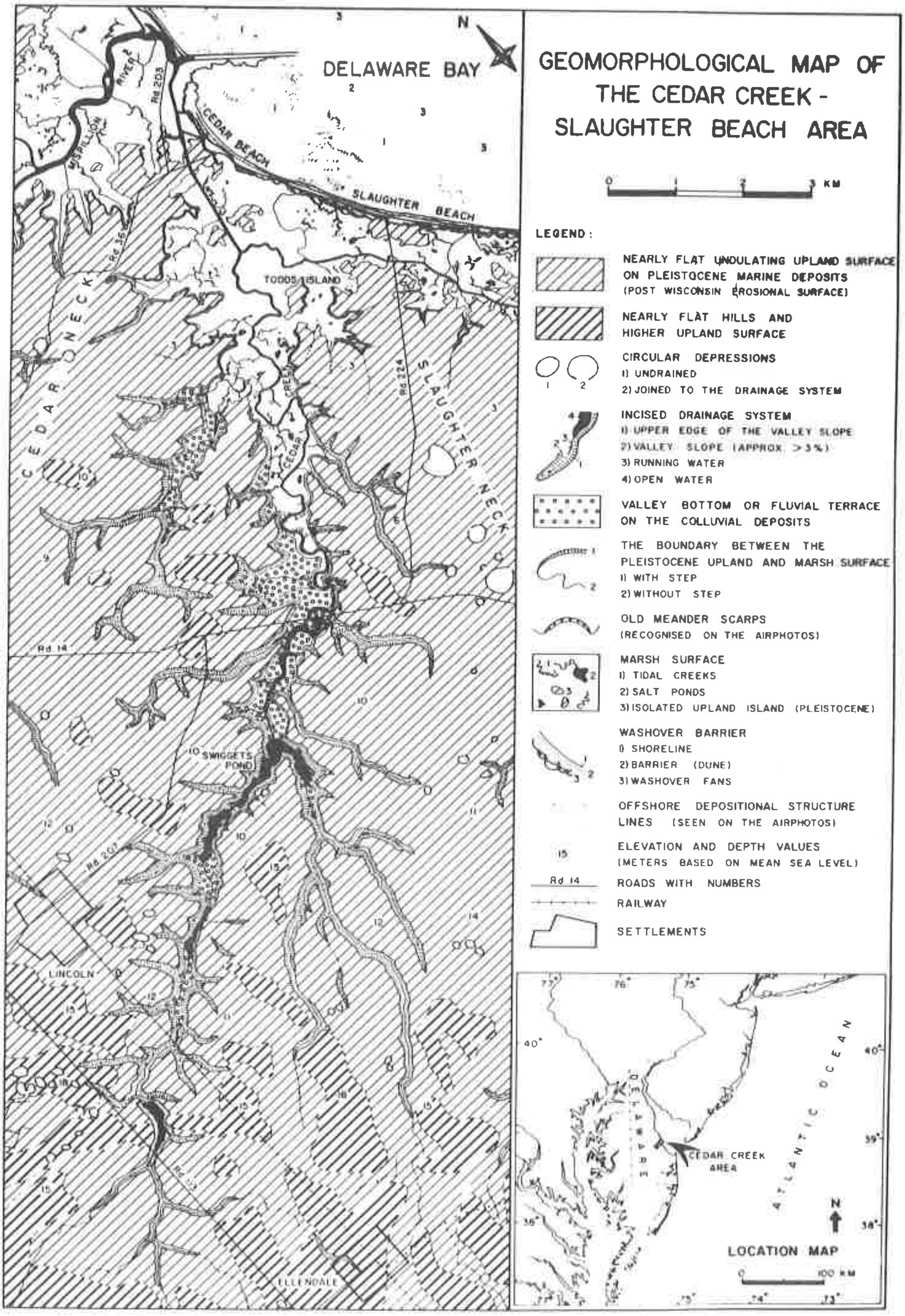


Figure 1. A geomorphological map of the Cedar Creek-Slaughter Beach area, including an index to the area of study.

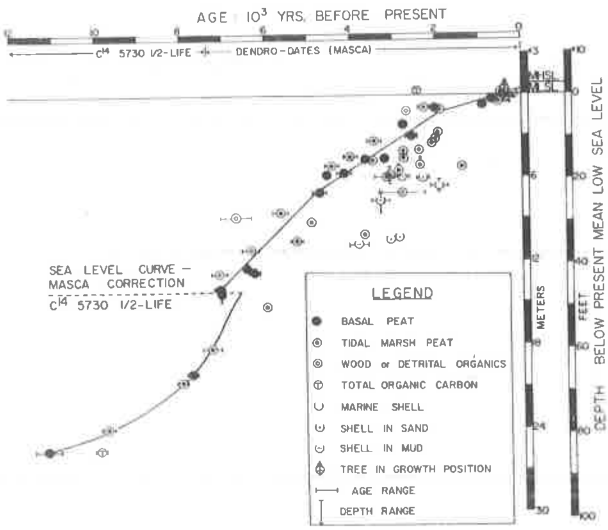


Figure 2. A local relative sea level rise curve based on radiocarbon dates from the Delaware coastal zone. Details of location and materials dated are published in Kraft, 1976a. Radiocarbon analyses in this study were corrected according to the MASCA correction tables (Ralph and others, 1973).

Detailed studies of the Pleistocene sediments of the Atlantic Coastal Plain were made by Jordan (1964). Jordon concluded that the Pleistocene Epoch deposits were formed in fluvial environments in the north and middle part of the State of Delaware and shoreline and nearshore-shallow marine sedimentary environments in the southern part of Delaware. This concept is now being questioned in the sense of specific ages of the depositional units. On the other hand, these sedimentary environments appear to be clearly defined. It is possible that the Quaternary sediments of the Atlantic Coastal Plain were deposited during high sea stands in the Sangamon and in mid-Wisconsin times. Regional aspects of Holocene sediments and the Delaware coastal area are published elsewhere (Kraft, 1971a; Kraft, 1971b; Kraft and John, 1976; and Weil, 1976).

METHODS

Stereographic airphoto analysis was made of the study area. Present fluvial and coastal environments were related to an incised drainage pattern on the older coastal plain (Figure 1). Sedimentologic and flora-faunal studies were made of the present coastal environments.

The results of these studies were then used as analogs for the interpretation of information obtained from the subsurface. Drilling in the area of study was by means of a truck mounted auger rig. The auger was carefully drilled into the subsurface in 6 m units and then lifted without spinning the auger. Thus subsurface samples were able to be obtained essentially "in-place" although with distortion of sedimentary structures. In addition, a number of cores were taken by pounding plastic pipe sections to depths of up to 3 m. Based on studies elsewhere (Kraft, 1971b, and Kraft and John, 1976), a great deal of information is known regarding the depth of incision of the Wisconsin valleys tributary to the ancestral Delaware River. In general, major tributary systems were incised to a depth of 30 m along the Delaware Bay and Delaware Atlantic shorelines approximately 14,000 years ago. Weil (1977) confirmed this deeply incised configuration of tributaries to the ancestral Delaware River in the Delaware Bay, by seismic means. Thus, this study presents an integration of analysis of surface morphology, presently ongoing erosion and deposition in the coastal zone, and subsurface evidences of the nature of coastal sedimentary environmental lithosomes of the Holocene.

THE GEOMORPHIC SETTING: CEDAR CREEK-SLAUGHTER BEACH AREA

Almost the entire coastal zone of Delaware Bay is surrounded by broad, low-lying salt marshes. When older coastal plain sediments of pre-Holocene age are encountered by the eroding shoreline, sand and gravel is supplied to the littoral transport system and small beaches occur. The shoreline is linear or arcuate, modified by prevailing wave patterns. The landward edge of present deposition of the coastal marshes is controlled by the highly irregular previously incised drainage system of Wisconsin time (Figures 1 and 3). The study area may be divided into four geomorphic areas: 1) the landward low-lying older coastal plain, 2) the wide presently forming coastal salt marshes, 3) the linear sand barrier, and 4) the adjacent tidal flats (Figure 1). The Pleistocene upland surface is believed to have formed in a shallow near-shore marine environment in Sangamon or mid-Wisconsin time. Elevations above sea level on this surface are generally 4-15 m, but in some areas reach 17-18 m in possible relict dune ridges or barriers formed during the last regression of the sea from the study area. At the edge of the present marsh-tidal creek environment, the slope of the upland surface increases sharply toward the marsh. This higher slope is part of the flank of deeply incised Wisconsin age valleys which have progressively infilled as the Holocene relative sea level rise continues (Figures 3 and 4 and Kraft and others, 1976). The upland surface is poorly drained and includes small circular depressions with a diameter

Figure 3. Aerial photo of the study area: Cedar Creek, Slaughter Beach, coastal washover barrier and salt marshes (U. S. Department of Agriculture). (1) The broad low-lying Spartina alterniflora marsh; (2) The washover barrier and village of Slaughter Beach; (3) A low "island" of Pleistocene sediments undergoing inundation; (4) Delaware Bay; (5) Pleistocene surface - coastal plain.

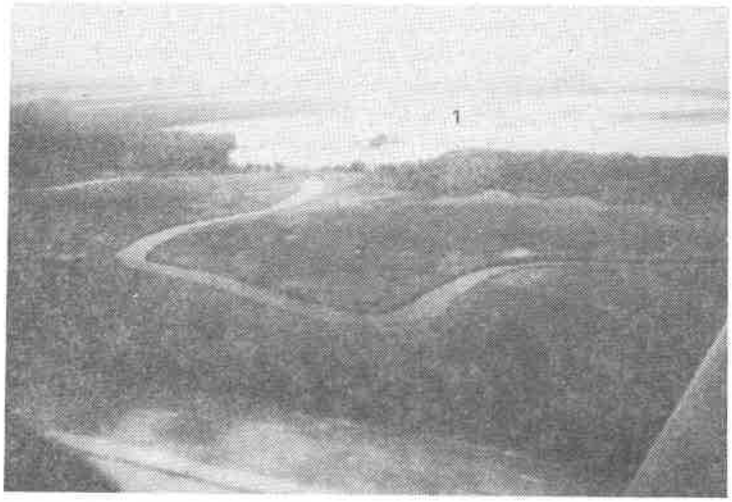


Figure 4. Central portion of the valley of Cedar Creek showing the relationship between the tidal marsh (2) and the Pleistocene Epoch higher surface (1).

of 100-200 m and depth less than 1 m (Figures 1 and 3). Possibly the circular forms are "Carolina bays", common on the Atlantic Coastal Plain from Maryland to Georgia. However, intensive farming has muted their features and their origins remain in question in Delaware. The Cedar Creek valley and associated salt marshes intrude landward along valleys incised in Wisconsin times. Present limits of tidal effects are at mill pond dams built in Colonial times. The upper portions of smaller streams such as Cedar Creek include coarser grained deposts derived from meandering and colluvial flow of sediments.

The floral character of the salt marsh surface is controlled by relative elevation of the tidal marsh. The lower tidal marsh is dominated by *Spartina alterniflora*. Toward the back of the barrier and the higher marsh areas near the transgressioned edge at the highland coastal plain surface, the marsh flora is dominated by *Distichlis spicata* and *Phragmites communis*. The marsh plants cover the marsh surface with a thick root system and form a baffle effect which traps finer size sediments at high tidal stages and tends to build the marsh surface upward in keeping with the gradual relative sea level rise in the area. In view of the highly irregular surface undergoing transgression, hills of pre-Holocene age are surrounded by marsh. These low-lying "islands" are covered with shrubs and trees. As relative sea level rises, these "islands" undergo a floral succession from shrubs and trees to high marsh floras.

The sandy estuarine washover barrier between the coastal salt marsh and Delaware Bay tends to maintain itself as it transgresses landward by coastal erosion across the low-lying salt marsh. Rates of erosion in the area of study are approximately 1 m per year but vary up to 8 m per year along the adjacent Delaware Bay shoreline to the north and south, as averaged from coastal survey maps over the past century (Kraft and others, 1976). Short-term rates of erosion are extremely difficult to predict. The area is subject to high intensity events such as hurricanes and northeasters. The low-lying coastal area has a tidal range of 1. 5 m and is therefore subject to storm overwash. Seaward of the washover barrier lies a 500 m wide organic muddy tidal flat. Short-term rates of erosion are difficult to determine since the mouth of the Mispillion River to the north is protected by two jetties, approximately 1. 5 km long and the direction of storm wave approach is highly varied.

The estuarine washover barrier between the coastal marshes and Delaware Bay is generally about 3 m above mean sea level. It varies in width 150-200 m from the bay to the limits of the sandy washover lobes. The washover barrier extends like a natural levee along the shoreline; thus, the small creeks that cross the marsh tend to flow parallel to the back of the barrier as tributaries to larger creeks, eventually draining into the larger Mispillion River. During times of storm erosion, the barrier is overwashed and massively eroded. The berm-beachface at the town of Slaughter Beach has varied over the past several decades from massive erosion and undermining of bulkheads to full wide beaches established by beach nourishment programs. Sand and gravel eroded from the beach is not replaced by the natural littoral transport system at the rate of storm loss. This may be caused by the jetties of the Mispillion River or by lack of source of additional course sediments to the littoral transport system. The inner tidal area in front of the barrier includes wide tidal flats of bay muds and sands and gravels eroded from the beach.

COASTAL PROCESSES

Climatic conditions in the area of study vary greatly on a yearly and long-term basis. In general, winter winds are stronger, with constant direction. In the Cedar Creek-Slaughter Beach area, the dominant wind direction is from land to sea along the coastline, which lies in a northwest-southeast direction. Thus the wind in opposition to the average direction of wave approach and may tend to decrease the power of waves as they impinge upon the shoreline in the area of study. However, winds and waves of hurricanes and northeasters are generally from the east. These storm waves are the most effective processes causing geomorphic change of the beaches in the study area. The fastest wind recorded in the past 50 year period along the Delaware Bay coastline was 130 km per hour. The average wave height in Delaware Bay is relatively low, 0. 5 m. The average tidal range in the area is

1.25 m with a spring tide range of 1.5 m. During extreme storms (50 year recurrence interval) a maximum effective wave height of 7 m with calculated extreme heights of 12 m can occur in Delaware Bay (Maurer and Wang, 1972). As a result of storm dominated wave activity the estuarine barriers remained low and are frequently overwashed.

Although the general littoral transport system in the Delaware Bay on the southwestern or Delaware shoreline is to the south, frequent exceptions occur. Variable wind directions in storms may reverse the littoral drift pattern. Milford Neck, lying to the north of the area of study provides a source of course sand sediment that flows southward into the area of study. However, much of this sediment is trapped by the jetties of the Mispillion River. These jetties cause a disruption of flow and accordingly a counter flow along Slaughter Beach in this area of study, providing for short periods of northerly littoral drift in the relatively protected area south of the jetties. Regardless, over the longer term, coastal erosion is much greater than coastal deposition. Accordingly the shoreline continues to move landwards and the marsh surface continues to build upward with sediment of the turbid waters of the flood tides flowing across the marsh surface. Based on an analysis of coastal processes impinging upon the study area, it is felt that the jetty of the Mispillion River, the resultant alteration in current flow patterns, extreme storm events, and continuing relative sea level rise are the prime reasons for exceptional rates of erosion and resultant landward migration in the Slaughter Beach area.

SUBSURFACE SEDIMENTARY UNITS IN THE CEDAR CREEK-SLAUGHTER BEACH AREA

Eleven holes were drilled in the study area. Radiocarbon analyses of five samples of salt marsh sediments believed to be in growth position were obtained. A number of interpretive cross-sections were made and are indexed on Figure 5. These cross-sections were constructed both parallel and perpendicular to the washover barrier. By means of these cross-sections and with a knowledge of relative sea level change in the area as provided by the radiocarbon dates in this study and elsewhere (Kraft, 1976a) paleogeographic reconstructions were made. The subsurface tests were drilled with a truck mounted auger rig. As 6 m of penetration were made, the augers were lifted without rotation. Thus the sedimentary sequences could be examined and sampled with reasonable confidence that the sediments studied were in proper stratigraphic position. However, sedimentary structures tended to be destroyed by this process. Test holes were drilled into the underlying Pleistocene sediments in an attempt to define the shape of the land surface during the peak of the Wisconsin glaciation. In general, the Pleistocene sediments are compacted sands and oxidized to yellow, orange, tan and

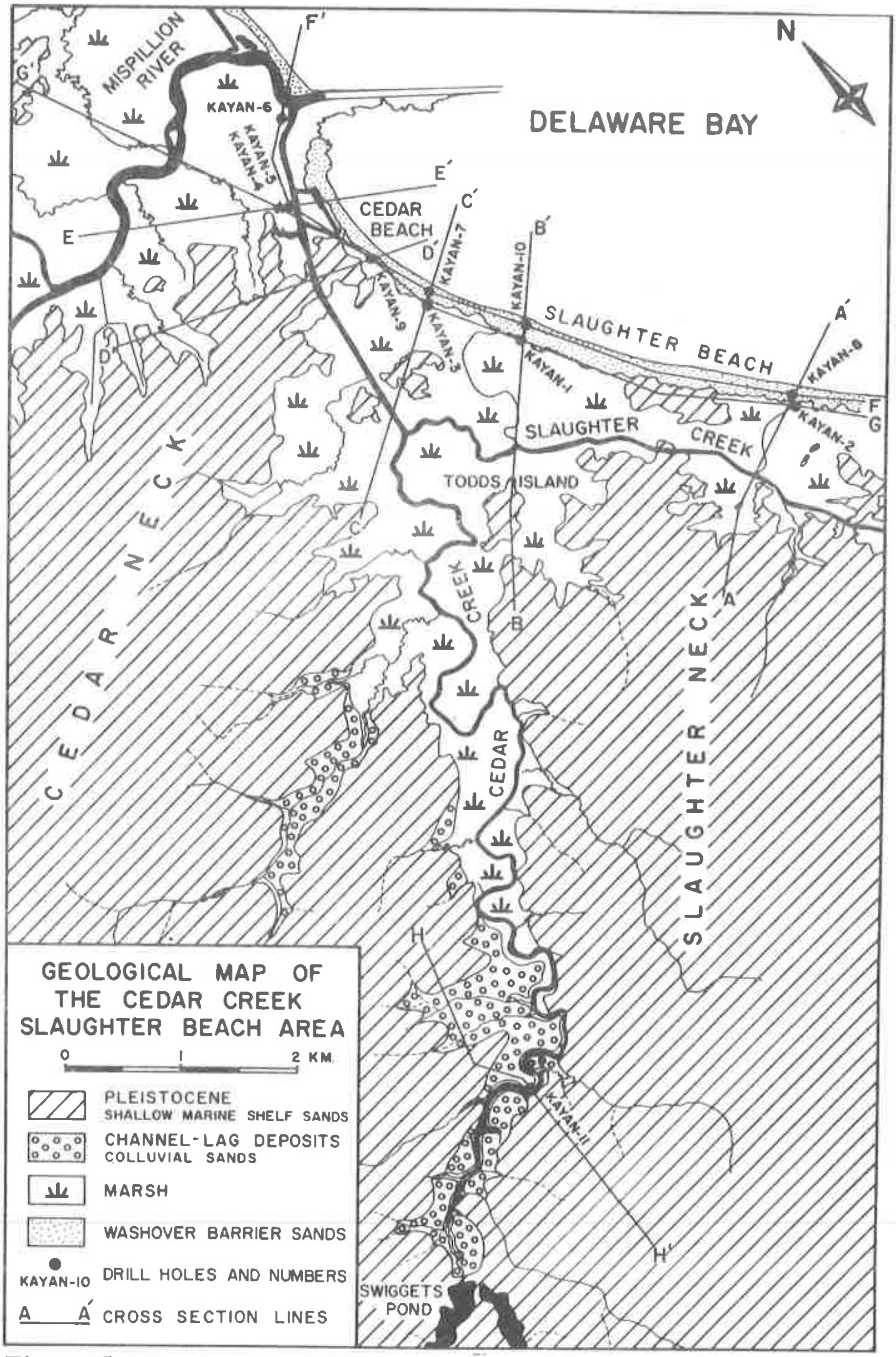


Figure 5. Geologic map of the Cedar Creek-Slaughter Beach area showing lines of interpretive subsurface cross sections.

light brown and light gray colors. No organic debris was found in the Pleistocene sediments. Depth to the Pleistocene surface is highly varied. Pleistocene sediments are exposed in the adjacent Coastal Plain at elevations up to 17-18 m above sea level. In the holes drilled for this study, the Pleistocene surface was found at depths up to 14 m below sea level. Further, in studies of adjacent areas along the entire coast of Delaware Bay and the Delaware Atlantic Ocean coast, Pleistocene erosional surfaces have been found at depths up to 35 m below sea level. Thus, the deeply incised erosion surface at the peak Wisconsin glaciation and resultant lower sea level had a total relief of approximately 50 m in the valleys tributary to the ancestral Delaware River (Kraft and John, 1976). The ancestral Delaware River flowed southeastward through central Delaware Bay into the area of the present Atlantic Ocean and thence eastward across the continental shelf. This, much larger river, was incised to over 65 m below present sea level. Thus, with knowledge of base levels of the major rivers tributary to the ancestral river, projections of depths of valleys can be made in the area of study. Stratigraphic sections interpreted and presented in the paper are based on a highly probable common base level of erosion of river valleys to the north and south during the peak Wisconsin Epoch.

Holocene fluvial sands occur as channel-lag deposits. These deposits usually consist of dark gray coarse sands with a low percentage of pebbles. They occur along the axes of Cedar Creek and its tributaries and can be easily differentiated from the Pleistocene sands and gravels by the fact that they are non-consolidated and not oxidized. These channel deposits occur at the base of the Holocene transgressive sequence.

The salt marsh deposits appear to keep pace with the continuing rise of sea relative to land. Thus the marsh surface itself may be used as a paleo-indicator of sea level stand if one can identify a marsh surface in the subsurface and date it by radiocarbon methods. Much of the marsh section is an organic clay with marsh plant debris scattered throughout. At times the marsh flora developed in relatively dense layers of organic sediment. Some of the sediments are clearly transported macerated debris. However, others are recognizable as marsh flora in growth position. This indicates a marsh surface and, if datable, is extremely useful in paleogeographic reconstructions. In some of the subsurface samples, lagoonal mud sequences are found. These lagoonal sediments consist of clay silts, generally gray or dark gray in color. A fauna of Elphidium clavatum and Crassostrea virginica Gmelin was found. These, plus a few rare ostracodes are a good indication of lagoonal conditions. Drill holes and trenches were also made through the coastal washover barrier deposits. Sediments of the washover barrier are viable, generally light in color, and moderately to poorly sorted with an average medium-coarse grained size. In local areas, well sorted, fine dune sands occur along the upper part of the barrier. However, in view of the frequent washovers that occur, dune fields are only a minor element of the barrier morphology. The beach-berm and washover deposits include a low percent of pebbles. The pebbles are quartz and silicified limestone. A number of these pebbles include fossil corals of Silurian and Devonian age, similar to those found in the Appalachian Mountains of northeastern Pennsylvania. These fossils are an evidence that the original source of sand and gravel in the Coastal Plain deposits of the Delaware region was erosion of paleozoic sediments in northeastern Pennsylvania. Presumably the sediment was transported to the lower Delaware Coastal Plain by glacial streams and redistributed and deposited by coastal processes during the Pleistocene Epoch.

Figures 6 to 13 show detailed cross-sections of the depositional environments found in the subsurface. The sediments are shown schematically. Environmental interpretations have been made based on direct correlation of lithologies found in the drill holes with sediments forming in present day coastal environments in the area of study. This has been further supported by comparison of Foraminifera, Ostracoda, and mollusks found in the drill holes, with those presently living in the coastal environments. Care was taken to determine whether or not marsh flora was transported redeposited debris or marsh flora in growth position. With careful study, it is possible to identify the marsh flora by megascopic and microscopic means; and therefore, determine whether or not a high or a low marsh flora was encountered (Allen, 1977). In addition, surface and subsurface environmental analyses elsewhere in the Delaware coastal region were used as supportive information in interpreting the subsurface environments in the Cedar Creek-Slaughter Beach area (Kraft, 1971a, Kraft, 1971b, Kraft and John, 1976, and Kraft and Others, 1976).

Figures 6 and 7 indicate the presence of a coastal back barrier lagoon surrounded by fringing salt marsh. These lagoonal conditions existed in some portions of the area of study from approximately 6000 years B. P. to 300-400 years B. P. The shapes of the Holocene subsurface sedimentary bodies are highly varied and in part determined by the pre-existing erosional topography of the land surface formed during the time of the peak Wisconsin glaciation. As sea level rose with the waning of the Wisconsin glaciation, coastal sedimentary environments migrated in a landward direction, infilling the valleys and keeping pace with rise of sea level through the Holocene Epoch. Thus the overall thickness of the Holocene sediments deposited has been determined by the rise in sea level over the transgressed land surface. Radiocarbon dates presented in the cross-sections allowed the development of a time frame for the deposition of these sediments. However, the actual shape of the sedimentary environmental lithosomes and resultant thickness of sediments of Holocene coastal environments is extremely varied, and dependent upon pre-existing erosional topography undergoing transgression. Cross-section Figures 8-10 are perpendicular to the barrier but do not include lagoonal muds. However, they show the highly

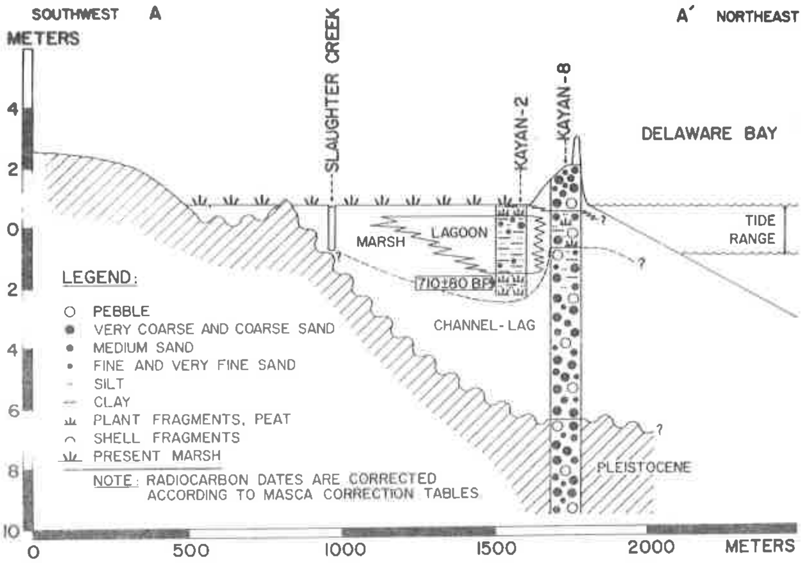


Figure 6. Geologic cross section perpendicular to the washover barrier shoreline.

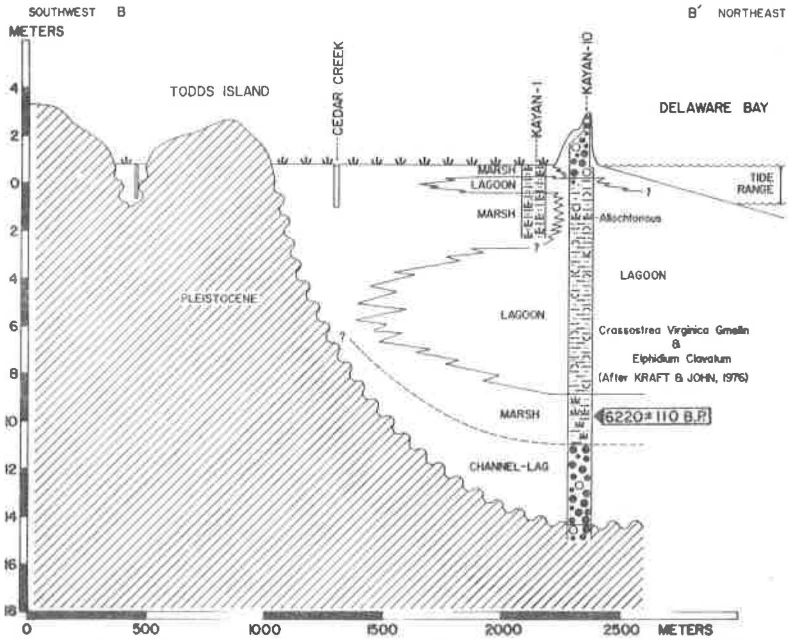


Figure 7. Geologic cross section perpendicular to the washover barrier shoreline. Radiocarbon and faunal data from Kraft and John is projected from 0.5 km to the southeast (see legend Figure 6).

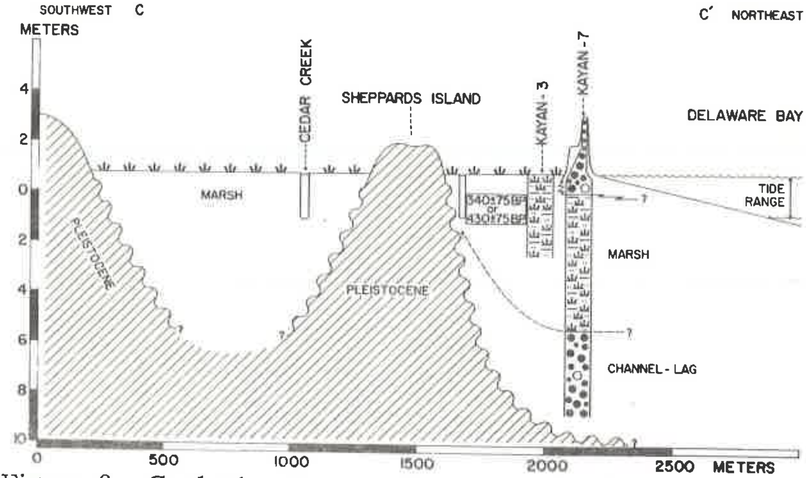


Figure 8. Geologic cross section perpendicular to the washover barrier shoreline (see legend Figure 6).

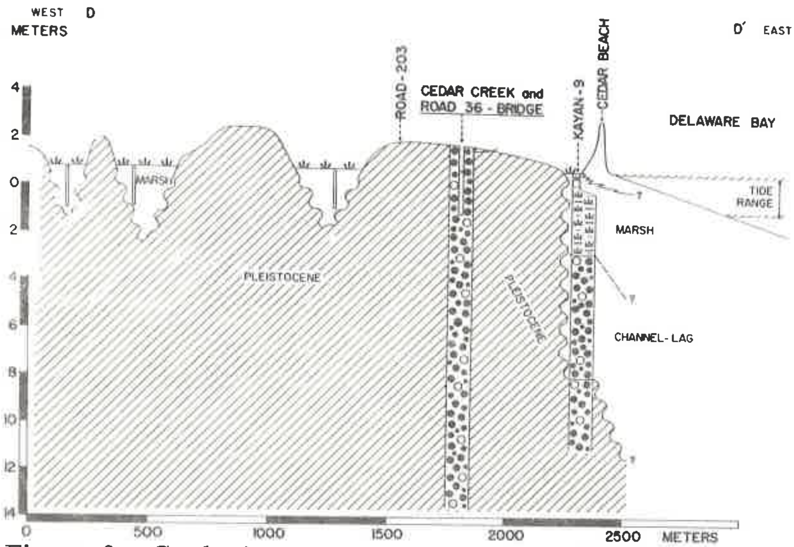


Figure 9. Geologic cross section perpendicular to the washover barrier shoreline (see legend Figure 6).

irregular nature of the pre-Holocene erosional surface. Presumably the areas covered by cross-section Figures 8-10 were in areas fringing coastal lagoons or further landward along the valleys of tidal creeks. Most important, Figures 6-10 show the highly irregular nature of the pre-Holocene erosion surface. Cross-section Figures 11 and 12 are

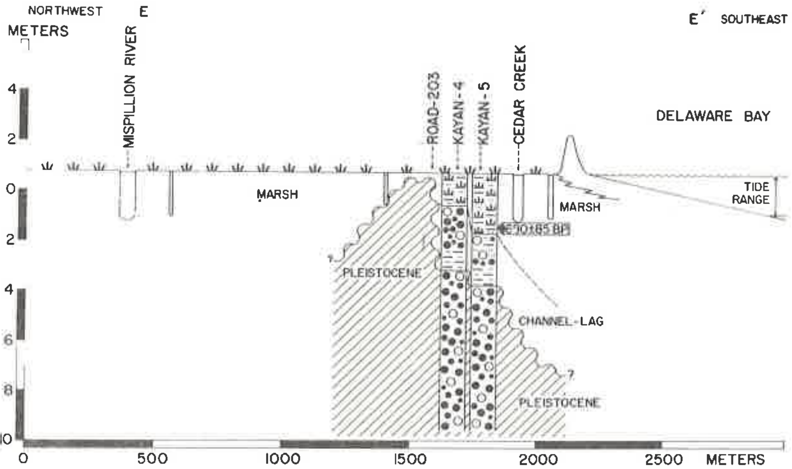


Figure 10. Geologic cross section perpendicular to the washover barrier shoreline (see legend Figure 6).

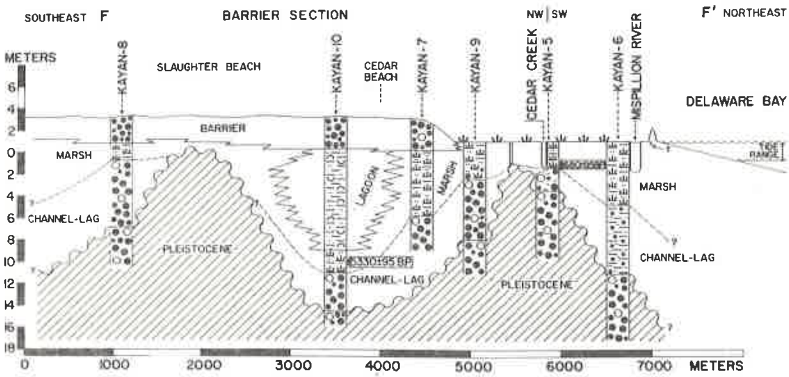


Figure 11. Geologic cross section parallel to and including a portion of the washover barrier shoreline (see legend Figure 6).

drawn roughly parallel to the coastal zone. They show a different perspective and display incised stream valleys that have been infilled with Holocene sediments. The oldest radiocarbon date of marsh flora in place found in the study area occurs in boring JCK-DH-71 dated at 6220$\pm$110 B. P. (Figure 7). Thus, this basal salt marsh sediment indicates a time of arrival of tidal waters into the Cedar Creek study area. Figure 11 is a cross-section parallel to the shoreline along the transgressive washover barrier. The barrier is transgressive landward over salt marsh and lagoonal environments and a highly irregular incised valley system that may be dated to early Holocene times in the

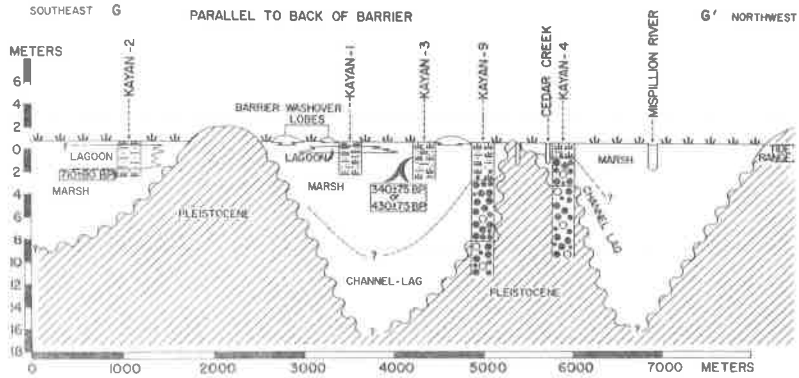


Figure 12. Geologic cross section parallel to and landward of the washover barrier cross section showing some of the barrier washover lobes (see legend Figure 6).

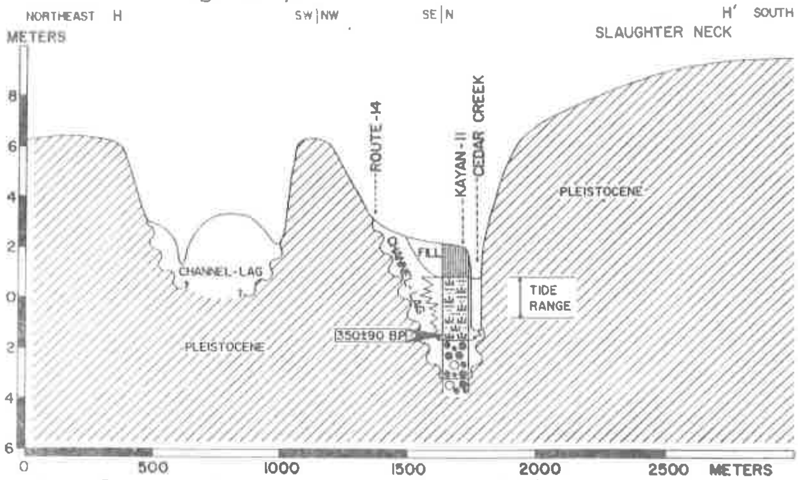


Figure 13. Geologic cross section of the upper Cedar Creek valley near the head of tidal penetration (see legend Figure 6).

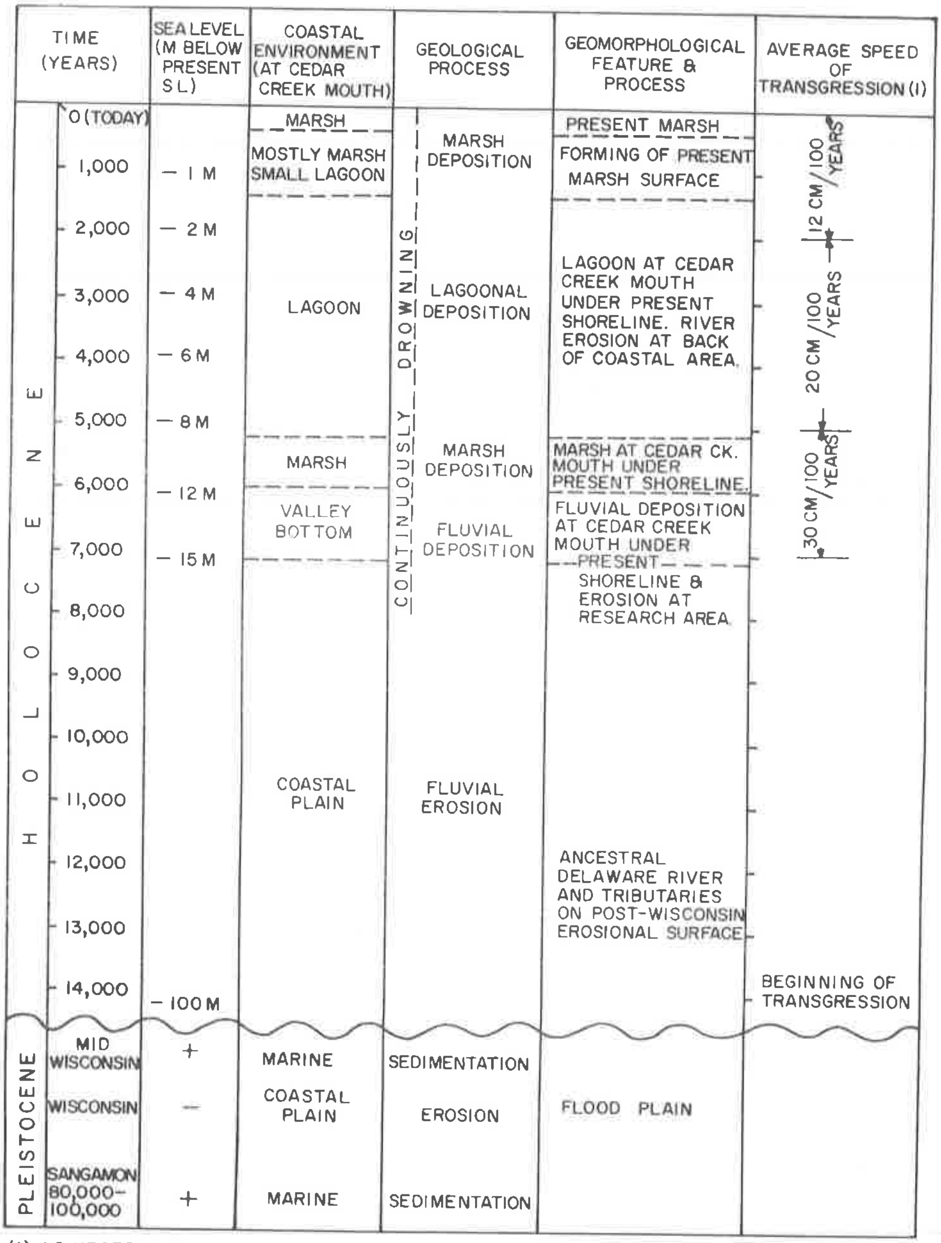
area of study but to the peak Wisconsin glaciation from the evidence elsewhere in the Delaware coastal zone. Cross-section Figure 12 is also parallel to the shoreline but landward of the barrier. Some of the washover lobes which extend landward on top of the marsh are shown. Cross-section Figure 13 shows the upper portion of tidal Cedar Creek. Here channel lag deposits are overlain by a tidal marsh sequence that infills the narrow, relatively deeply incised upper valley of the ancestral Cedar Creek.

PALEOGEOGRAPHIC RECONSTRUCTION

Figure 2 shows a relative sea level curve for the Delaware coastal area based on data from Belknap and Kraft (1977) and Kraft (1976a) in the Delaware coastal zone along Delaware Bay and the Atlantic coast. As may be seen from Figure 2, there are many variable interpretations that could be made from this data. Kraft (1976a) chose to use only basal organic rich sediments (only 30% organics, not a true peat) in drawing the sea level curve. Information regarding present day coastal processes and sedimentary environments, surface and subsurface sedimentology, present and past geomorphologies, and rates of sea level change relative to land have been presented in cross-sections Figures 6-13 and Figure 14. These data may be used to make paleogeomorphological reconstructions in the study area. Two paleogeomorphological interpretations of the coastal zone at 2000 and 5000 years before present are shown on Figure 15. Sea levels used in determining these maps were based on the local relative sea level rise curve for the Delaware coast (Figure 2) plus the information gained from radiocarbon dated samples from this study area.

Five thousand years before present, the Pleistocene upland area extended beyond the present Cedar Creek-Slaughter Beach area. At that time, the Holocene transgression had created a coastal lagoon near the mouth of the ancestral Cedar Creek and the Slaughter Beach area. Evidence for the existence of this coastal lagoon was found by Kraft and John (1976) approximately 0.5 km southeast of drill hole Kayan-10 (B-B, Figure 7 and F-F, Figure 11). Other back barrier lagoons in the coastal zone of lower Delaware Bay were identified by Elliott, who noted a very large lagoon in the area of the present Lewes Creek marsh north of Lewes, Delaware (Elliott, 1972).

In view of the fact that the Holocene transgression occurred across a deeply incised surface, marsh fringes around the mid-late-Holocene lagoons were relatively narrow. With a lessening of the rate of relative sea level rise in late-Holocene time, salt marshes began to spread over wider areas. Based on the configurations of sedimentary environmental lithosomes shown in cross-section Figures 6-13, a sequence of depositional events and paleogeographic construction may be envisaged. Fluvial sediments were probably accumulating in the channels of Cedar Creek and its tributaries throughout the Holocene Epoch. At 2000 years before present (Figure 15) the lagoon was still present and the barrier shoreline had eroded and migrated at least 1 km. Sea level was approximately 2 m below present at that time. The marsh of 2000 years ago was much wider as the land surface being overridden at that time had a lower relief. The deeply incised valleys were already infilled. Fluvial deposits were still forming along the heads of the small creeks running into the lagoon and Delaware Bay. As the marshes spread and the relief between the marsh surface and the upland surface became less, thinner deposits of channel lag gravels formed at the



(1) J.C. KRAFT 1971a, D.F. BELKNAP 1975, J.C. KRAFT 1976a, 1976b.

Figure 14. A summary of geomorphological features and sedimentary environments related to the Holocene transgression and geological processes active in the area of study: Cedar Creek-Slaughter Beach coastal zone of Delaware Bay.

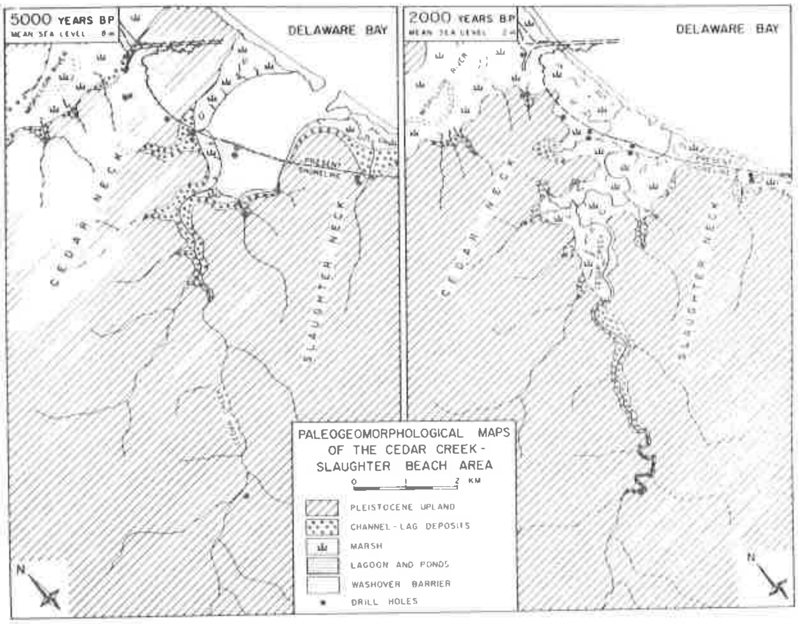


Figure 15. Paleogeographic reconstruction of the Cedar Creek-Slaughter Beach area for 2000 years before present and 5000 years before present.

heads of the valleys.

Comparisons of the two reconstructions show that the lagoon existed in the area for a relatively long time. Radiocarbon dates of the marsh that eventually encroached across this lagoon are relatively young, 300 to 400 years B. P. Similar conclusions were reached regarding other ancient lagoons along the Delaware Bay coast (Kraft, 1971b). It is possible that the shallow coastal back barrier lagoons were present when European settlers first reached North America. However, the radiocarbon evidence tends to suggest that the lagoons closed about 400 years B. P. A gradually narrowing marsh surface extends up the valley of Cedar Creek. As relative sea level continues to rise, this marsh surface will build landward and upward and tend to cover the channel lag deposits of the center of the Creek bed and further override the low-lying valley edges. In general, the drainage patterns of mid-Holocene time were the same as present; however, the barrier lay further seaward and the marsh lagoon sequence was below present sea level. Although there is evidence of abandoned marsh channels in some of the salt marsh areas, they are not common. Historical maps of the study area and adjacent coastal marshes to the north and south show no evidence of migration of meanders. Accordingly, it is hypothesized that the meander loops of the tidal creeks have remained in a relatively stable position for a long time.

CONCLUSIONS

Marine environments first intruded into the Cedar Creek valley area in the vicinity of Slaughter Beach about 6200 years B. P. From then until now as sea level gradually rose, the coast eroded and the estuarine washover barrier migrated landward and upward across the coastal salt marsh. At the same time, the salt marsh surface rose at a rate similar to the relative sea level rise. At one time, a large coastal lagoon occurred in the area of Slaughter Beach. This lagoon was finally filled and covered by a salt marsh approximately 300-400 B. P. Today the Slaughter Beach barrier continues to undergo intensive erosion and washover and a slow migration landward. Without needed beach nourishment by means of dredging sands and gravels from offshore shoals, the barrier would gradually disappear within the next century and a more marshy shoreline would take its place. With continued relative sea level rise and resultant rise of the marsh surface, the late Holocene transgression will expand landward as salt marsh conditions continue to encroach upon the low-lying coastal plain. The summary of geomorphological and geological processes of coastal environmental change presented in Figure 14 will probably continue into the short term geological future. Thus man would be well advised to adapt to these ever changing shoreline environmental configurations in planning future construction or development of the coastal zone.

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THE PORT DEPOSIT GNEISS REVISITED

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ABSTRACT

Geologic mapping in Cecil County, Maryland by Higgins (1972) led to the recognition of supracrustal rocks within the belt of rocks mapped as Port Deposit Gneiss by earlier workers and interpreted by them as wholly plutonic. Remapping along the Harford County side of the Susquehanna River reveals that the Port Deposit Gneiss as formerly delineated consists of 5 mappable subdivisions: (a) diamictite; (b) gneissic biotite quartz diorite; (c) quartz augen gneiss; (d) coarse-grained biotite-quartz-plagioclase gneiss; and (e) massive porphyritic leucogneiss. Unit (a) is a metasedimentary rock. Units (b) and (d) are plutonic igneous rocks in the form of sheets or thick sills. Unit (e) is metavolcanic rock composed of massive dacitic flows and tuff sheets. Unit (c) is of equivocal origin but is interpreted as a metaplutonic rock, probably originally trondhjemite.

INTRODUCTION

Recent work in the Appalachian Piedmont by Crowley (1976), Seiders and others (1975), Pavlides and others (1974), and Higgins (1972) has provoked lively interest in and controversy about a group of granitoid gneiss bodies that Hopson (1964) earlier had interpreted as syntectonic plutons. Prominent in this debate are various interpretations of the Port Deposit Gneiss, a body of granitoid gneiss in northeastern Maryland that has been controversial since the days of Grimsley (1894) and Bascom (1902). The recent exchange between Seiders (1976) and Higgins (1976) underscores the need for yet another look at this complex, oft-studied, and poorly understood body of rock.

The Port Deposit Gneiss is typical of a group of similar rocks that includes the Norbeck, Kensington, and Relay masses in Maryland (Hershey, 1937; Hopson, 1960, 1964; Southwick, 1969; Higgins, 1972) and the Occuquan and Dale City masses in Virginia (Seiders and others, 1975). All of these bodies are, broadly speaking, in the granodiorite to quartz diorite range of composition, of granitoid textural appearance, metamorphically foliated, and in contact with one or more subunits of the Wissahickon Formation or the James Run Formation and its equivalents (see Higgins, 1972, and Seiders and others, 1975 for details). Moreover, all of them have radiometric U-Th-Pb ages in the neighborhood of 500-560 million years (Higgins, 1972, table 1, p. 1010-1012 and references therein; Seiders and others, 1975). If these rocks are true plutons that are intrusive into their wallrocks, which is how they were interpreted prior to 1972, their radiometric age marks a minimum age for the Wissahickon and James Run sequences. If, on the other hand, these rocks are pseudogranitic metasediments and metavolcanics, as Higgins (1972) argued, their age becomes more nearly a direct measure of the depositional age of the enclosing supracrustal formations. Higgins (1972) based his interpretation largely on the structures, textures, and chemical composition of the Port Deposit Gneiss, with some corroborative evidence from other localities. Seiders (1976) and Seiders and others (1975) argue for the true plutonic interpretation, mainly using evidence from the Occoquan and Dale City masses in Virginia but with supplemental observations on the Port Deposit Gneiss.

Acknowledgments

Reviewed by W. P. Crowley and G. W. Fisher. Financial support provided by Maryland Geological Survey and Macalester College. I extend special thanks to M. W. Higgins for a preprint of the geologic map of Cecil County, Maryland, by Higgins and Conant, and for stimulating discussion and debate.

PSEUDOGRANITIC VERSUS PLUTONIC ROCKS IN THE PORT DEPOSIT GNEISS

Prior to 1960 the Port Deposit Gneiss (Granodiorite of Hershey, 1937) was thought to be entirely plutonic. In 1960 Hopson recognized that the inclusion-rich granitoid rock exposed just below Conowingo Dam, interpreted earlier by Hershey (1937) as a xenolith-laden contact phase of the Port Deposit "granodiorite", is not an igneous rock at all, but is a metasedimentary slump deposit. This material is similar in all respects to the Sykesville and Laurel Formations of south-central Maryland, which were first thought to be intrusive rocks also (Williams 1895), but which Hopson (1964) clearly demonstrated to be of sedimentary origin. Most of the distinctive slump deposit rock (diamictite) in northeastern Maryland has been mapped as a facies of the Wissahickon Formation (Fisher and Higgins, 1971). The stratigraphic position of the Conowingo Dam occurrence is not certain, however, and Higgins (1972) has included those rocks in the informally defined Conowingo gneiss.

Although Hopson (1960) recognized diamictite near Conowingo

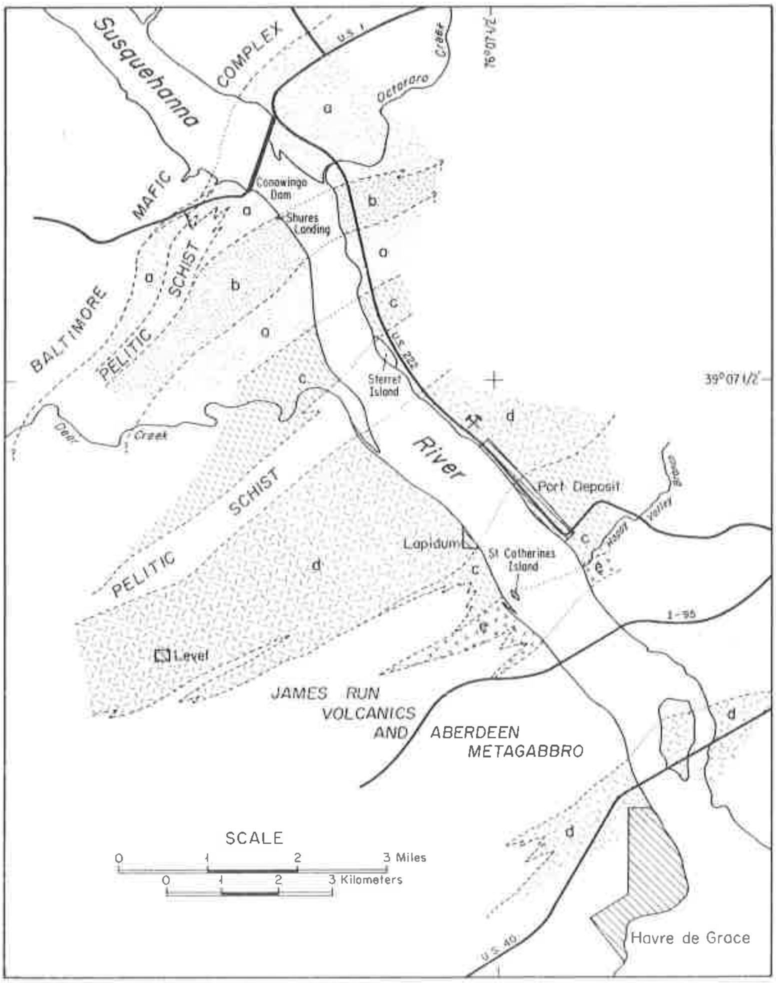

Figure 1. Geologic sketch map showing the distribution of rock types along the Susquehanna River within the belt mapped as Port Deposit Gneiss (or Granodiorite) by previous workers. Five kinds of granitoid felsic gneiss, labelled (a) - (e), are recognized and are described under the corresponding letter in the text. Map modified from Southwick and Owens (1968), Higgins (1972), and Higgins and Conant (in press) on basis of field studies in 1975.

Dam, he concluded that it was invaded by true plutonic rock south of Octoraro Creek (Figure 1). He thought the pluton was syntectonic, "...with magmatic flowage passing into protoclastic granulation, in turn passing indistinguishably into metamorphic deformation and recrystallization as the mass became sufficiently competent to transmit regional deforming stresses. " (Hopson, 1960, p. 30-31).

A second rock type that is grossly plutonic in appearance, but actually is not plutonic, is massive metadacite and/or metadacite tuff belonging to the James Run Formation. Most of the James Run, which flanks the Port Deposit Gneiss on the southeast, consists of metamorphosed dacitic to basaltic flows, tuffs, and volcaniclastic rocks that are well stratified and contain vestiges of primary volcanic structures in less metamorphosed areas (Southwick, 1969, p. 45-59; Higgins, 1972, p. 1001-1002). Part of the section, however, consists of thick, massive units of former dacite or rhyodacite porphyry, some of which may have been tuff originally and some of which may have been massive flows. Where sufficiently metamorphosed those rocks may closely resemble gneissic quartz diorite or granodiorite (Southwick, 1969, p. 64), and have been mistaken for them in parts of Cecil and Harford Counties, Maryland (Higgins, 1972, p. 1002).

Near the center of the Port Deposit belt as mapped by Hershey (1937) are large outcrops of coarse grained, thoroughly recrystallized, even-textured biotite-quartz-plagioclase gneiss of overall plutonic appearance. Despite the fact that virtually all traces of plutonic texture have been destroyed by metamorphic recrystallization and the development of two foliations (Southwick, 1969, p. 122, fig. 24; Higgins, 1976, p. 1527, fig. 3), the rock is so uniform in composition that everyone who has studied it concurs on its plutonic parentage. This rock occurs in the large quarry north of Port Deposit that has become the informal type locality for Port Deposit Gneiss, and also is well exposed between Lapidum and Rock Run in Harford County.

Thus, there is absolutely no doubt that the belt of rocks mapped by Hershey (1937), and termed by him the Port Deposit Granodiorite Complex, is in fact polygenetic. There are metasedimentary rocks (diamictite) in the northern part, metaplutonic rocks in the middle, at and near the Port Deposit quarry, and metavolcanic rocks in the southern part. Substantial parts of the mass are difficult to decipher in terms of origin, however, and it is probable that agreement will never be reached as to the exact proportions of plutonic, volcanic, and sedimentary rocks that were present initially.

DESCRIPTION OF GRANITOID ROCKS WITHIN THE PORT DEPOSIT GNEISS

Within the area shown as Port Deposit Granodiorite on the geologic map of Hershey (1937, plate 20) there are at least five units of granite-like rock that can be recognized along the Susquehanna River. These are shown on Figure 1 as units (a) through (e). All of these rock

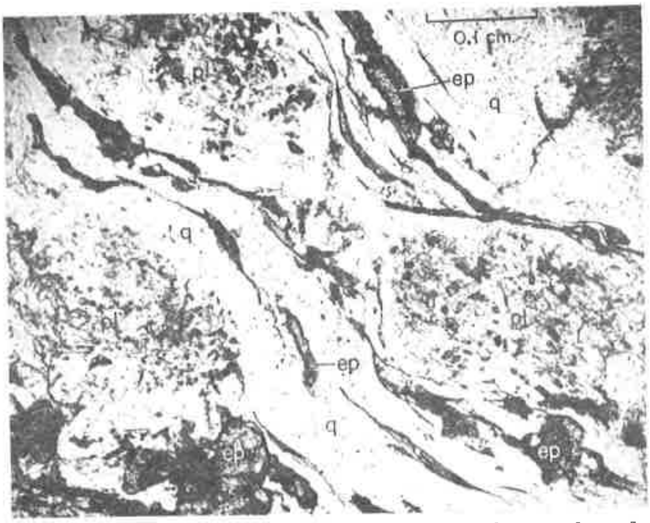

Figure 2. Photomicrograph of extensively sheared and recrystallized gneiss from unit c of the Port Deposit Gneiss. pl = saussuritized plagioclase; q = quartz; ep = epidote. Fine-grained white mica is abundant.

types have been affected by multiple deformation and metamorphism. In places two foliations are well developed; locally there has been very extensive metamorphic shearing accompanied by recrystallization of abundant secondary quartz, Na-plagioclase, clinozoisite, and muscovite (Figure 2).

(a): Pebble- and Boulder-Bearing Diamictite

This striking rock, thoroughly described by Hopson (1960), is best exposed in waterworn outcrops just below Conowingo Dam. The bulk of the rock is a weakly foliated aggregate of quartz, plagioclase, and mica that superficially resembles quartz diorite; its sedimentary parentage is indicated by its high quartz content (locally more than 50 percent), relict sand grains and granules, and the presence of scattered boulder-size clasts, many of which are best interpreted as intrabasinal in origin (Hopson, 1960). The same lithology also occurs in a 1-1.5 km-wide belt about 2.5 km south of Conowingo Dam, (incorrectly mapped as sheared quartz diorite by Southwick and Owens, 1968) and in small lenses closely associated with the stringer and lenses of metasedimentary schist that all previous workers have mapped within the greater outcrop area of the Port Deposit.

(b): Medium-Grained Gneissic Biotite Quartz Diorite

Somewhat metamorphosed biotite quartz diorite, locally with hornblende also, occurs in a belt about 1 km wide south of Shures Landing in Harford County. It contains less quartz and muscovite, and more plagioclase and biotite, than the quartz augen gneiss (unit c) which it otherwise resembles closely. Moreover, it possesses well-preserved hypidiomorphic granular texture with subhedral zoned plagioclase, wedge-shaped volumes of interstitial quartz (Figure 3), and scattered euhedral crystals of zoned allanite (Figure 4). The southern contact of this material against diamictite is indistinct and difficult to map precisely with the available exposure. The biotite quartz diorite here described was not recognized by Higgins (1972), but was recognized by Hopson (1960 and pers. comm.).

(c): Quartz Augen Gneiss

Oval grains of blue-gray quartz on the order of 5-10 mm long constitute about 20-25% of this rock; the remainder is plagioclase, biotite, finer-grained quartz, and muscovite together with variable small amounts of iron-poor epidote and garnet. Foliation ranges from strong to barely detectable. In well foliated rocks particularly the quartz grains resemble clastic granules; in less foliated rocks the hand specimen appearance is more "granitic". Higgins (1972) mapped a broad belt of this material roughly between Octoraro Creek and Sterret Island and interpreted it as finer-grained, granule-bearing diamictite, gradational to the boulder-bearing variety south of Conowingo Dam. He interpreted very similar rocks in and south of the village of Port Deposit as a transitional igneous facies between plutonic rocks (on northwest) and volcanic rocks of the James Run Formation (on southeast). I concluded earlier (Southwick, 1969, p. 64) that most of this material is of plutonic origin and still believe that much of it is (Figure 1). Evidence for this conclusion is developed in separate sections to follow.

(d): Coarse-Grained, Highly Foliated Biotite-Quartz-Plagioclase Gneiss

A thoroughly recrystallized, even textured, granitoid gneiss with grains as large as 2 cm crops out in the center of the Port Deposit belt, at and near the Port Deposit quarry, and continues southwestward at least as far as Level. Strong deformation and recrystallization have obliterated original textures, but its coarse grain-size and the presence of aplite dikes indicate that the original rock was plutonic quartz diorite and granodiorite. Quartz, plagioclase, biotite, and variable small amounts of microcline are major minerals of probable igneous origin; muscovite, chlorite, epidote, and minor garnet have developed during metamorphism. The amount of modal quartz ranges from 27-45 percent.

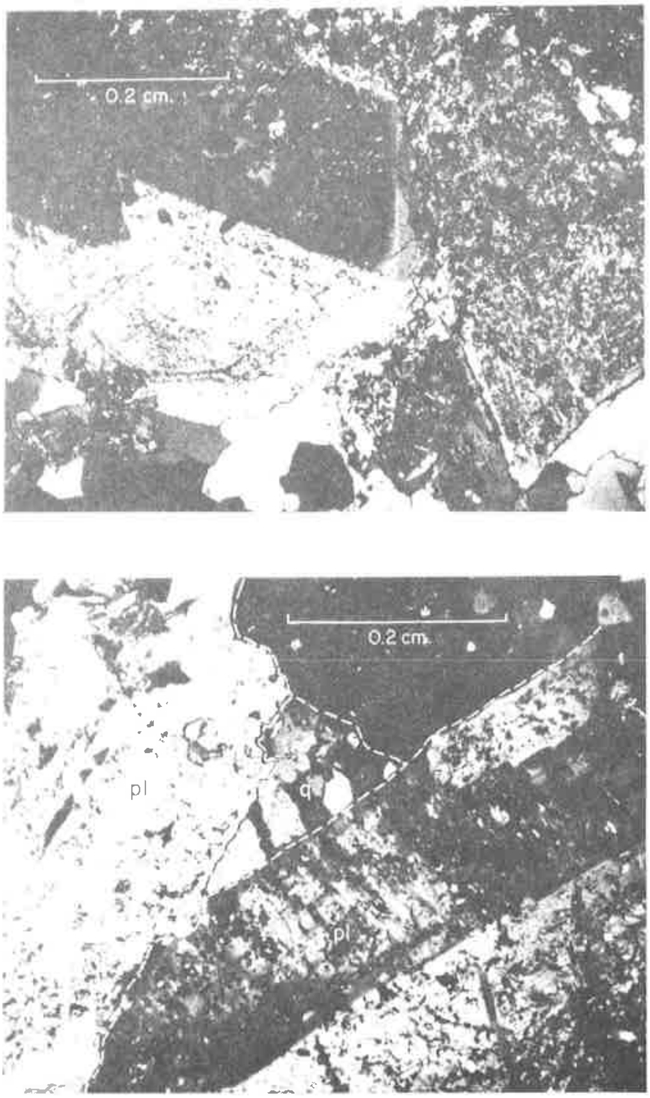

Figure 3. a. (top) Photomicrograph of zoned, subhedral plagioclase (saussuritized) in medium-grained biotite quartz biorite (unit b) of the Port Deposit Gneiss. b. (bottom) Photomicrograph of wedge-shaped interstitial quartz (q) between subhedral plagioclase crystals in unit b of the Port Deposit Gneiss.

Near Lapidum in Harford County this coarse gneiss surrounds numerous slabs of fine- to medium-grained amphibolite. These slabs

Figure 4. Photomicrograph of euhedral, compositionally zoned allanite crystal in unit b of the Port Deposit Gneiss.

range in thickness from 0.1 to 4 m, are oriented parallel to the foliation of the gneiss, and occur generally in groups of two to six or more in which the slabs are spaced several cm to several m apart. The largest have directly observable strike lengths of several tens of meters. The gneiss between the slabs is identical in all respects to that well away from them. Although some of the amphibolite slabs may be mafic dikes that were emplaced relatively late in the history of the area, but prior to the latest deformation, some of them clearly are mafic inclusions. Behind the houses in Lapidum is a 2 m-thick slab that is abruptly truncated by coarse granitoid gneiss. The foliation of the gneiss wraps around the end of the slab in a manner suggesting flow foliation of partly crystalline magma around a solid xenolith (Figure 5). The relationships corroborate Hopson's (1960) conclusion that magmatic flowage preceded protoclasis and solid-stage metamorphism during syntectonic emplacement of a pluton. Higgins (1972, p. 1000-1001) has suggested that this part of the Port Deposit may have been a very shallow, surface-breaking pluton analogous to the Tatoosh pluton in Mt. Ranier Park (Fiske and others, 1963).

(e): Massive Porphyritic Leucogneiss

South of Happy Valley Branch in Cecil County are large cliffs composed of very light-colored porphyritic gneiss. Plagioclase phenocrysts on the order of 5 mm in size are set in a weakly to strongly foliated groundmass of quartz and feldspar. Megascopically the foliation

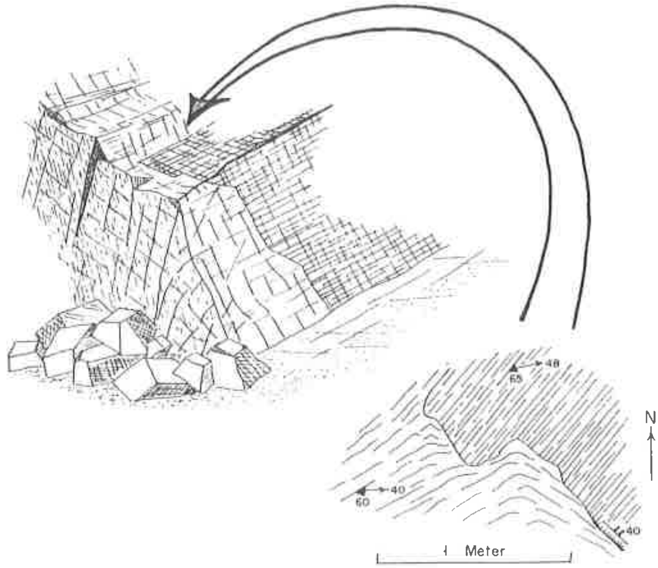

Figure 5. Field sketch of outcrop (vegetation removed) about 8 meters NW of Lapidum Road, about 100 m WSW of corner in village of Lapidum. Face of low bluff behind small clearing is a slab of medium- to fine-grained, well lineated amphibolite; behind and at the southwest end of the amphibolite slab is coarse, well foliated biotite-quartz-plagioclase gneiss of the Port Deposit (unit d). A long-abandoned small quarry in the gneiss is just southwest of the clearing. Foliation in the gneiss near the end of the slab is discordant to that in the amphibolite (see insert); it becomes contorted and deflects around the amphibolite. The outcrop is densely overgrown with honeysuckle and poison ivy.

is carried by polycrystalline blebs of biotite, chlorite, and epidote. The rock is not strikingly different in outcrop characteristics from other plutonic or possibly plutonic rocks of the Port Deposit belt, and earlier workers (Bascom, 1902; Grimsley, 1894; Hershey, 1937; Southwick, 1969) construed it as plutonic. On the basis of its textures and gradational contacts with well stratified tuffaceous rocks of the James Run Formation, Higgins (1972, p. 1001-1002) concluded that the leucogneiss was volcanic, probably representing a thick pile of rhyolite, rhyolite tuff, and dacite tuff, and mapped it with the James Run Formation. Higgins' assignment of these rocks is correct. The rocks between St. Catherine's Island and Havre de Grace in Harford County are chiefly felsic volcanic rocks, not plutonic as shown on the Harford

County geologic map (Southwick and Owens, 1968). They appear to be intruded by thin sheets of former diabase and gabbro, probably related to the Aberdeen Gabbro to the southwest, and also by thin sheets of granitoid gneiss, in part porphyritic, that may be related to larger plutonic masses within the Port Deposit belt.

CHEMISTRY OF THE GRANITIC GNEISSES, WITH SPECIAL REFERENCE TO THE PORT DEPOSIT

Chemical criteria were used by Hopson (1964) to help distinguish pseudogranitic diamictite from true plutonic rocks in Howard and Montgomery Counties, Maryland. He argued that the diamictite is too silica- and alumina-rich for an igneous rock. To illustrate this point he plotted norms for diamictite in the normative Q-Ab-Or-H_2O diagram (Tuttle and Bowen, 1958) and found that they all fell within the primary quartz field. Norms of undoubted igneous rocks, on the other hand, plotted mainly within the primary feldspar field. Hopson (1964) noted further that the diamictite contains normative corundum, whereas most fresh igneous rocks do not. He attributed the excess silica and alumina in diamictite to the well-known geochemical processes that lead to enrichment of these substances in the sedimentary cycle.

Most chemical analyses of Port Deposit Gneiss (broadly defined) are high in normative quartz and contain some corundum (Southwick, 1969; Higgins, 1972; see norms, table 1). Higgins has followed Hopson's (1964) chemical reasoning and used the high silica and alumina content of the Port Deposit as additional evidence that much of it was of sedimentary derivation.

These chemical criteria for sedimentary origin may be questioned on several grounds. First, because the Port Deposit Gneiss contains about 15 percent normative An, the five-component system Ab-An-Or-Q-H_2O is a better model for interpreting its petrochemistry than the anorthite-free haplogranite system studied by Tuttle and Bowen (1958). The more general five-component system has not been fully worked out in the laboratory, but enough experimental data are available to construct an approximate but reasonably well constrained phase diagram for certain geologically realistic conditions (Presnall and Bateman, 1973). Figure 6 is an exploded view of the liquidus of the system Ab-An-Or-Q-H_2O, water saturated at about 5kb, as deduced by Presnall and Bateman (1973, p. 3186) and used by them to interpret the crystallization history of the Sierra Nevada batholith. The average of five analyses of Port Deposit Gneiss reported by Southwick (1969) is plotted as point PD, which falls within the primary phase volume of plagioclase and comfortably above the quartz-plagioclase boundary surface. This point projects into the quartz field on the Q-Ab-Or face, but its actual 3-deminsional position within the tetrahedron is in the plagioclase

Table 1. Average Normative Compositions*.

	1		2		3	4	5
	σ	Avg	σ	Avg	Avg		Avg
Q	5.84	39.42	5.90	40.80	38.69	29.23	24.92
Or	2.97	17.10	3.61	5.22	12.12	9.45	14.57
Ab	5.40	19.57	6.19	33.57	27.51	17.76	37.04
An	7.28	7.12	6.04	12.59	14.57	22.41	14.63
C	3.43	4.96	1.01	.88	.15**	---	1.63
Wo	---	---	---	---	.20**	2.32	---
En	.89	3.95	1.48	1.38	8.71	1.50	
Fs	1.04	2.69	1.52	2.55	2.69	5.16	2.59
Il	.32	1.14	.17	.58	.56	.87	.55
Mt	.59	2.81	.98	1.32	1.48	2.90	1.69
Ap	.25	.40	.21	.19	.21	.28	.12
Number of samples		9		13	2	1	2

Note: 1, diamictite from several localities in Maryland and Virginia, including occurrence at Conowingo Dam; 2, felsic layers in James Run Formation and related units; 3, coarse biotite-quartz-plagioclase gneiss, unit d, Port Deposit; 4, gneissic hornblende-biotite quartz diorite, unit b, Port Deposit; 5, quartz augen gneiss of debatable origin, unit c, Port Deposit.

*Compiled from data previously published by Hopson (1964), Southwick (1969), Higgins (1972), Southwick and others (1971), and Seiders and others (1975).

**One analysis C-normative, the others Wo-normative.

volume. Moreover, the modal analyses of subunits b, c, and d (Table 1, columns 3, 4, 5) all plot in the same general portion of the diagram.

The phase relations at 5kb suggest that a magma having the composition of the Port Deposit Gneiss could, despite its high silica content, first crystallize plagioclase followed by K-feldspar and quartz. Remnant textures from the biotite quartz diorite belt of the Port Deposit (unit b) indicate exactly that crystallization sequence (Figure 3).

Other experimental work in synthetic granite systems, reviewed by Wyllie and others (1976), clearly shows that quartz is the primary liquidus phase at pressures greater than about 6kb in granitic melts containing about 2 percent water. If more than 2 percent water is present, quartz appears at lower pressures (down to about 3kb at 5% H_2O). Under these conditions early-formed quartz would tend to stay in

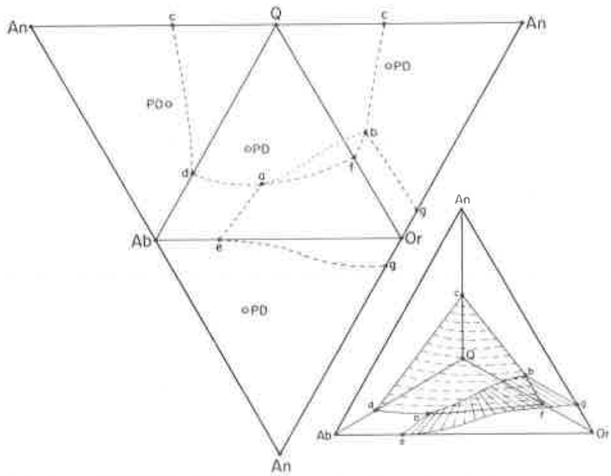

Figure 6. Exploded view of the liquidus of the system Ab-An-Or-Q-H_2O, saturated, at about 5 kb, as deduced by Presnall and Bateman (1973). The average of available analyses of units b, c, and d of the Port Deposit Gneiss plots at point PD.

suspension in the magma and might even float toward the top of magma chambers owing to its low density. Anomalously quartz-rich igneous rocks conceivably could result from such quartz flotation; these would be expected to grade into rocks with more normal silica contents.

Thus, the composition of the Port Deposit Gneiss could yield the paragenetic sequence plagioclase-K-Feldspar-quartz if crystallization occurred in the neighborhood of 5kb total pressure. The sequence quartz-plagioclase-K-feldspar might occur if pressure were 6kb or more, of if the magma were anomalously wet. There are scraps of textural evidence from the biotite quartz diorite belt substantiating the more common plagioclase-first paragenesis; it is possible, on the other hand, that some of the rocks with quartz augen actually contain early quartz phenocrysts.

Clearly the foregoing arguments do not prove that the Port Deposit Gneiss is entirely magmatic; indeed, Higgins (1972) is quite correct in pointing out that part of the complex is, in fact, metasedimentary. The important point is that quartz-rich gneiss in the Port Deposit is not excluded from plutonic parentage by its rather silica-rich composition.

A second objection to the chemical argument used by Higgins (1972) lies in the very real possibility that the composition of Port Deposit rocks has been modified during metamorphism. Careful studies by many geologists have found significant differences between fresh and metamorphosed volcanic rocks in the proportions of alkalis, CaO,

Al_2O_3, and SiO_2 (see, for example, Anderson, 1968; Hughes, 1972; Gunn, 1975). There is considerable likelihood that changes of a similar (but not necessarily identical) nature accompany the metamorphism of plutonic rocks, especially when pervasive foliation develops to allow free circulation of solutions.

Textural evidence in the Port Deposit Gneiss indicates that iron-poor epidote and muscovite have grown metamorphically at the expense of intermediate plagioclase and K-feldspar, respectively, with concomitant production of albite and quartz. Substituting iron-free clinozoisite for epidote to simplify schematic reactions, the observed mineralogical relationships may be written as:

$$3KAlSi_3O_8 + H_2O \longrightarrow KAl_3Si_3O_{10}(OH)_2 + 6\,SiO_2 + K_2O$$

$$3\text{ K-feldspar} + H_2O \longrightarrow 1\text{ muscovite} + 6\text{ quartz} + K_2O$$

and

$$4\,CaAl_2Si_2O_8 + H_2O \longrightarrow 2\,Ca_2Al_3Si_3O_{12}(OH) + Al_2O_3 + 2\,SiO_2$$

$$4\text{ Anorthite} + H_2O \longrightarrow 2\text{ clinozoisite} + Al_2O_3 + 2\text{ Quartz.}$$

To the extent that K^+ is removed from the system during metamorphism and deformation, these reactions will produce a bulk enrichment in SiO_2 and Al_2O_3. This will enhance the contents of normative quartz and corundum, and cause any igneous rock which undergoes such metasomatism to acquire metasediment-like chemical characteristics.

The foregoing metamorphic argument obviously does not prove that the Port Deposit rocks necessarily experienced metasomatic alkali loss. It is offered merely as a plausible mechanism by which a chemically normal plutonic igneous rock might be modified toward the corundum-normative, quartz-rich composition actually observed.

A final point concerning the chemistry of these rocks has to do with corundum in the norms. As shown in Table 1, true diamictites from near Conowingo Dam and elsewhere in the Maryland - Virginia Piedmont contain substantially more normative corundum than do undoubted metavolcanic rocks from the James Run Formation and other related metavolcanic units. The corundum content of Port Deposit Gneiss (other than unit a) is in the same range as the metavolcanics, suggesting that the alumina enrichment in both is either a metamorphic phenomenon, and inherent magmatic characteristic, or some combination of the two.

Cawthorn and Brown (1976) have pointed out that many calc-alkaline intrusive and extrusive suites show a trend from diopside - to corundum-normative compositions with increasing SiO_2 content. Thus it may well be that the modest corundum contents of James Run felsite and Port Deposit Gneiss are a carryover from original composition, perhaps modified somewhat by metasomatism.

THE PROBLEM OF THE QUARTZ "EYES"

Pea-size augen of blue-gray quartz occur widely in the northern part of the Port Deposit Gneiss (units a, b, and c of Figure 1), and have been interpreted by Higgins (1972) as relict sedimentary granules in more or less even-textured diamictite. Some of these are granules (in unit a), but there are reasons to suggest an igneous origin for others and to interpret the quartz augen gneiss of unit c as metamorphosed trondhjemite.

Experimental work reviewed by Wyllie and others (1976) clearly shows that quartz may be an early-formed mineral in granitic melts that crystallize at elevated pressure. Early quartz should form more or less equant grains in plutonic rocks, in contrast to the highly irregular, amoeboid form that late-stage quartz typically adopts. Early quartz phenocrysts, when subjected to metamorphic shearing, would tend to become spindle-shaped and also polycrystalline, thus transforming into the so-called quartz eyes.

Although examples of unquestionable plutonic igneous rocks with quartz augen are not numerous, there are at least two such occurrences. The Lower Precambrian Saganaga Tonalite of northeastern Minnesota (Goldich and others, 1972; Hanson, 1972) contains conspicuous quartz "eyes" that are on the order of 1 cm in length and are interlocking aggregates of 1-2 mm crystals. Quartz also is interstitial to subhedral, antiperthitic oligoclase, the only other essential mineral in the rock. The eyes have been interpreted as partly resorbed quartz phenocrysts by Hanson and Goldich (1972, p. 188). They are visible not only in the central part of the Sanganaga batholith, where the tonalite is virtually unaltered and unstrained, but can be recognized easily toward the margins where the rock has been sheared, metasomatically altered, and the augen have been flattened to length-width ratios as great as 7:1. The Saganaga Tonalite is less silica-rich than the Port Deposit (21-25% modal quartz vs. 25-40%; 55-67% SiO_2 vs. 61-75%) and has undergone a quite different sort of synkinematic alteration. The two are comparable, nevertheless, in that they both are composed chiefly of sodic feldspar and quartz, are leucocratic, and are associated with arc-type metavolcanic rocks of basalt to dacite composition. The Saganaga pluton is an early syntectonic intrusion that was unroofed before the end of tectonic activity and contributed detritus to polymict but dominantly volcanogenic sedimentary rocks higher in the section (Ojakangas, 1972 a, 1972 b).

The Middle Precambrian trondhjemite near Rio Brazos, New Mexico (Barker and others, 1974) contains ovoid to subhedral polycrystalline quartz eyes as large as 20 mm in length. The eyes comprise about 5 percent of the rock. Barker and others (1974, p. 705-706) give the following petrographic description: "...Some of the trondhjemite consists of quartz eyes and blocky, subhedral plagioclase (calcic to median albite) grains set in a fine-grained, wholly recrystallized groundmass; the remainder shows a seriate, inequigranular fabric. The trondhjemite is foliated and has been subjected to metamorphism of lower amphibolite facies. Blue-green hornblende, epidote, biotite, chlorite, microcline, and white mica are other stable phases."

Average chemical and modal compositions of the Rio Brazos trondhjemite (Barker and others, 1974, p. 706; 1976, p. 191) are very similar to the most silicic of the Port Deposit rocks. A norm calculated from the average analysis of the Rio Brazos contains 40.9 percent quartz and 2.79 percent corundum. Some samples of the Rio Brazos contain as much as 50 percent normative quartz (Barker and others, 1976, p. 193). Furthermore, the quartz eye in the Rio Brazos photographed by Barker and others (1974, p. 706) is strikingly like the quartz eyes in the Port Deposit figured by Higgins (1972, p. 997).

CONCLUSIONS

On the basis of the foregoing chemical and textural arguments I conclude that not all of the quartz-rich augen gneisses in the Port Deposit are necessarily metasediments. Half or more are plutonic igneous rocks with trondhjemitic affinities. High-silica, high Na/K rocks chemically and modally similar to the Port Deposit, and covering a similar compositional range, are found in association with calc-alkaline volcanic suites of all ages, and appear to be an important component in the igneous evolution of convergent tectonic zones.

Regardless of the total proportion of plutonic rocks in the Port Deposit belt, the mass seems not be be a deep-rooted batholith as formerly interpreted (Southwick and Owens, 1968, cross sections). Geologic mapping by Higgins in Cecil County (Higgins and Conant, <u>in press</u>) and by Crowley (1976) in Baltimore County has revealed synformal closures on both ends of the Port Deposit terrane. This, together with Bromery's (1968) conclusion from geophysical evidence that mafic rocks underlie most of the Port Deposit Gneiss at relatively shallow depths, lends considerable support to Higgins' view (1972) of a tabular, sheet-like form for the Port Deposit Gneiss (<u>sensu vasto</u>). It appears that the Port Deposit is a series of inter-tonguing sill-like granodiorite to trondhjemite sheets, felsic volcanic rocks, and associated sedimentary materials. About two-thirds of this section consists of plutonic sheets, which lie between and in places include septa of supracrustal rocks. The effects of metamorphism and multiple deformation make it difficult to distinguish plutonic from volcanic from sedimentary components with certainty in all places, and it is likely that controversy over detail will continue. However, the following generalizations seem valid:

(1) The terrane southeast of St. Catherines Island and Happy Valley Branch is chiefly volcanic and/or subvolcanic, consisting of felsic porphyries that probably represent flows, tuffs, and shallow sills.

(2) The terrane between St. Catherines Island and Shures

Landing is largely plutonic granodiorite, tonalite, and trondhjemite, but with large inter-sill septa of supracrustal rocks, including diamictite.

(3) The terrane northwest of Shures Landing and the mouth of Octoraro Creek is chiefly diamictite. The diamictite appears to widen to the northeast in Cecil County, perhaps because of tectonic thickening as it nears the axial zone of the regional synform.

(4) The entire complex of sills, volcanic rocks and sedimentary rocks has stratigraphic and chemical affinities to the James Run Formation. The James Run is in part stratigraphically beneath the Port Deposit, but the two may interfinger to a large degree and therefore may be interpreted as nearly contemporaneous. The Port Deposit - James Run Complex may comprise the deeply eroded remnants of a mature volcanic island arc, a suggestion first made informally by Clifford Hopson about 15 years ago and developed further, but in somewhat modified form, by Crowley (1976).

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FURTHER OBSERVATIONS ON GLACIATION IN THE SHINING ROCK AREA, NORTH CAROLINA

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ABSTRACT

No evidence has been found for mountain glaciation in the Shining Rock quadrangle of North Carolina. U-shaped valleys do not exist, but V-shaped or open-pan shaped valleys occur. Valleys that lead from large ravines lack the eroded step-like profiles or bedrock depressions common to glacial terraines. Heads of the largest and steepest ravines do not have full-bodied cirque forms. Moraines have not been found anywhere across the region. Striated, polished or faceted clasts are absent from colluvium or regolith. Till is absent along creek bottoms, road cuts or valley sides. Bedrock outcrops now exposed on mountain tops and in saddles bear no glacial striae, grooves or other abrasional marks. The region lacks streamline-molded topographic forms. Most cross-valley ridges extend to valley floors and show no truncation. Stream valleys lack deposits of relict outwash gravels. Melt-water channels cutting across bedrock ridges are conspicuously absent. Flanks of high summits are not mantled by glacially transported boulders and are not horn-shaped. Boulder talus deposits along stream valleys are abundant and suggest a periglacial not a glacial environment during Wisconsin time. It would seem that even during the Wisconsin maximum, snowbanks seldom persisted into the summer season.

INTRODUCTION

Incipient relict glacial features in the Shining Rock and Sam Knob quadrangles of North Carolina were outlined by Haselton (1973). The 1973 report was written hoping it would stimulate other investigators to more carefully examine field areas across the crest of the Southern Appalachians in an effort to confirm or deny mountain glaciation.

Few field investigators believe Wisconsin glaciation occurred in this region, however, there are some who claim (J. Harrington, oral communication) that deposits of till and cirque-forms can be found at elevations below 3,500 feet (1067 m). Loren Raymond (1977) reports he has evidence for cirque-like features on the northwest and east sides of Grandfather Mountain in North Carolina.

After further field work, the writer has failed to find any conclusive proof for Wisconsin mountain glaciation even at altitudes of 6,200 feet (1890 m).

Acknowledgments

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DISCUSSION

There are several field criteria that do not support the thesis of alpine glaciation in the Shining Rock area:

1. Major and tributary valleys are either pan-shaped or V-shaped. There are no U-shaped valleys that typify glaciated areas. (Figures 5, 6, 7 and 8).

2. In the field, when observing principal valleys in the downstream direction, one notes the absence of glacially eroded step-like profiles or rock basins.

3. Most valleys at the heads of the steepest and deepest ravines lack the clear, deep amphitheater shapes of true cirques but do show some cirque-like morphology. For example, note the valley-head profiles of Daniel Ridge and Right Fork Creeks (Figure 1).

4. Valley cross sections immediately downstream from the heads of ravines are V-shaped. Please refer to Right Fork, Cove Creek and Daniel Ridge Creek valleys on Figure 1 and refer to Figures 5 and 6.

5. No moraines have been found across valleys immediately downstream from the deepest ravines nor do they occur across the principal valleys of the region; Davidson River and the East Fork of the Pigeon River.

6. Striated, polished, or faceted boulders are entirely absent throughout the area. Where striated or polished bedrock surfaces are found, they are the result of faulting.

7. High up in what some would assume to have been source areas for glacial ice, stream cobbles lack glacial markings.

8. With the exception of eroded areas or steep bluffs, the tops of the highest peaks are mantled by a rather well developed soil with surprisingly little coarse material (Figure 3). What appear to be boulders in Figure 3 are tree stumps. Saprolite and colluvium 4-6 feet (1.2-1.8 m) deep cover valley sides and the crests of some ridges

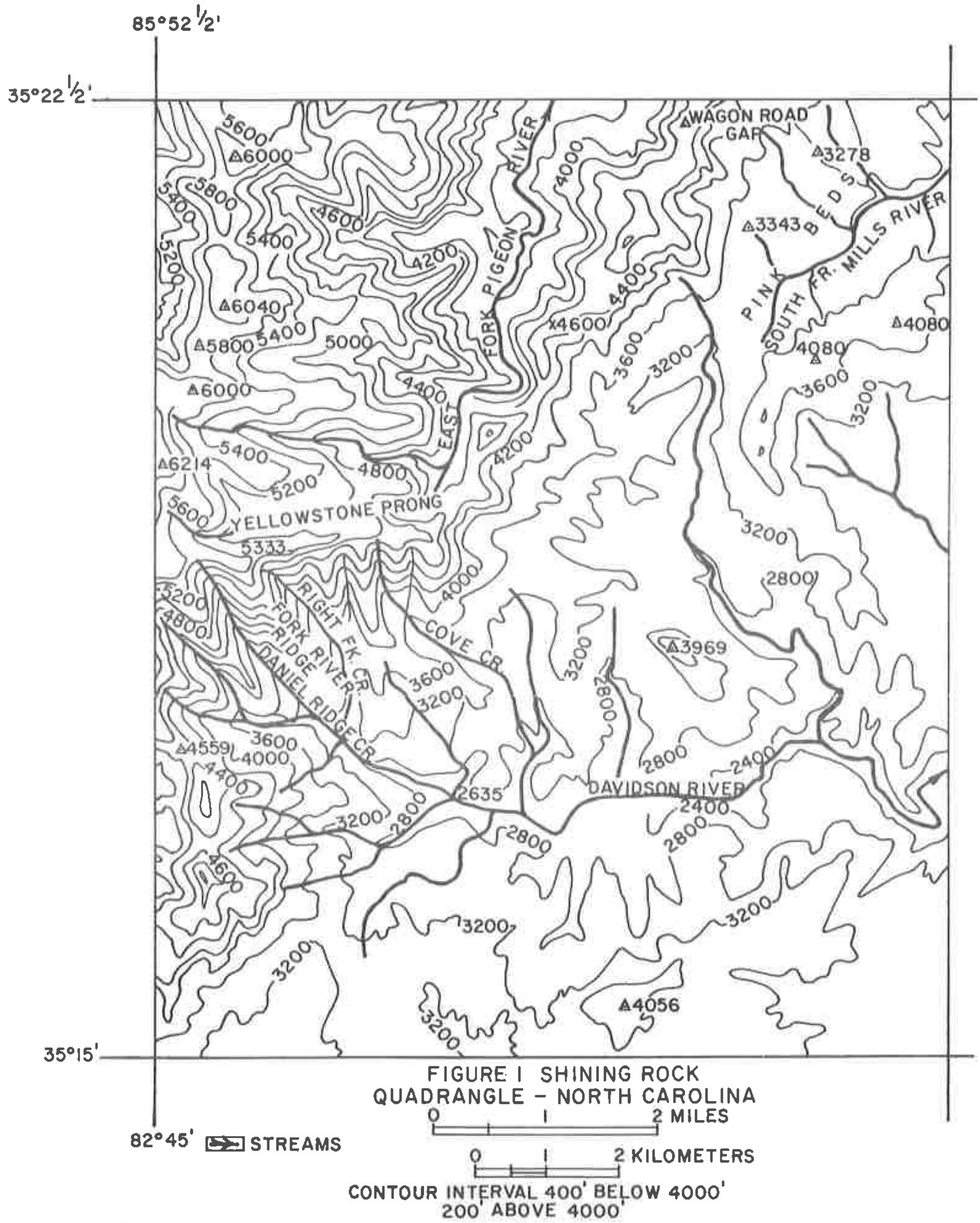

Figure 1. Location map, showing west half of Shining Rock Quadrangle, North Carolina.

(Figure 4) at elevations above 4, 400 feet (1341 m) and contain no striated cobbles or glacially modified clasts.

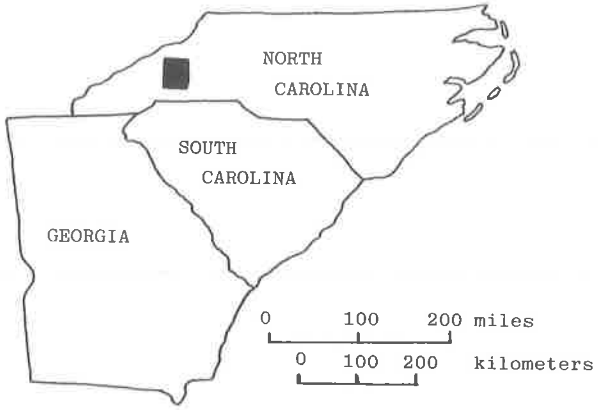

Figure 2. Regional map showing the location of the field work as seen in the dark square.

9. At the location called the Pink Beds (Figure 1), an appropriate area for any glacial ice to have formed a piedmont lobe, there is no till, no morainal deposits, or any sign of glacial modification. Since the floor of this broad valley is above 3,000 feet (915 m), if glaciation had occurred in these mountains in late Wisconsin time, one might expect to find some relict ice-erosional or depositional forms on this broad valley floor.

10. In all reaches of this quadrangle there is a complete lack of streamline molded forms namely, stoss and lee topography, whalebacks or roche moutonnees.

11. Side ridges entering all major valleys and principal tributaries extend unbroken to the valley floors (Figure 5). They seldom exhibit the abrupt truncated ends, referred to as truncated spurs, that characterize ridges in glaciated areas. The exceptions being where ridge truncations result from stream undercutting, strong joint control and mass wasting or faulting. Even in the upper reaches of the highest valleys spur trimming is noticeably absent (Figure 6).

12. Certainly any Wisconsin glaciation would have stripped off most of the colluvium or colluvial blanket thus providing abundant detritus which would have been washed down these valleys creating deep valley outwash fills. Such fills would now be trenched and terraced. Such high level residual gravel deposits are missing along major valleys such as the East Fork of the Pigeon River or the Davidson River (Figures 1 and 7). There are some low, poorly developed terraces composed of modern floodplain deposits but they do not demonstrate the necessary glacial meltwater conditions under which overloaded braided channels normally develop. Furthermore, gravel deposits along valley sides should also have developed as Kame terraces or other ice-contact deposit forms, and none have been found.

Figure 3. Broad, gently sloping summit peak 6214 (1895 m) as seen on west side of Figure 1. View is looking north. What appears to be boulders are tree stumps. There is an absence of a boulder cover. Trail in upper right cuts through soil to bedrock.

Figure 4. Angular bouldery colluvium resting on saprolite just west of the Shining Rock quadrangle at an altitude of 5,000 feet (1524 m). This is the kind of material that some call "till".

Figure 5. Interlocking spurs (ridges) looking west from the summit of peak 6,214 (1895 m), Shining Rock Quadrangle. Upper portion of this valley is above 4,400 feet (1341 m) and is V-shaped. No ridge ends are glacially truncated.

Figure 6. V-shaped valley with overlapping cross-valley ridges that have not been truncated by alpine glaciation. This is the first valley north of Yellowstone Prong (Dark Prong) seen on Figure 1. View is east from peak 6214 (1895 m). Lower part of valley in this view is above 4,800 feet (1463 m). Note the absence of bedrock basins along the valley floor.

Figure 7. The upper reaches of the East Fork of the Pigeon River Valley. The narrow boulder-choked channel can just be seen through the trees. Valley walls have scattered patches of boulder talus. Valley bottom has no remnant of outwash fill. Elevation of channel floor in this view is 4,200 feet (1280 m). Note V-shaped across profile of valley.

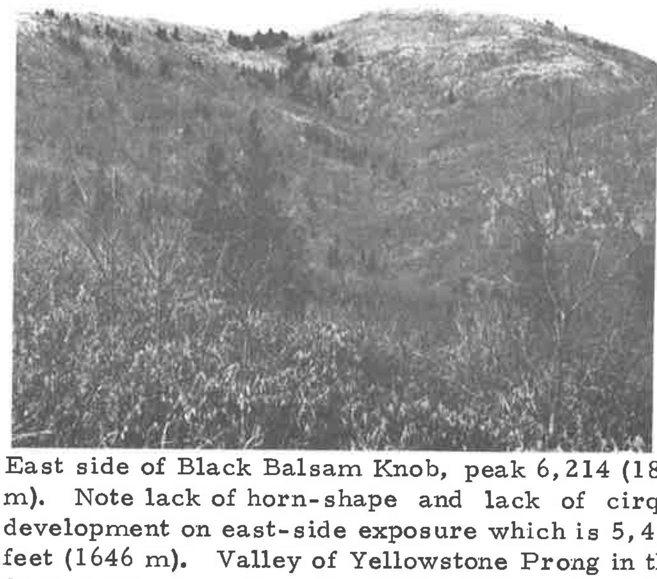

Figure 8. East side of Black Balsam Knob, peak 6,214 (1895 m). Note lack of horn-shape and lack of cirque development on east-side exposure which is 5,400 feet (1646 m). Valley of Yellowstone Prong in the foreground.

13. One finds no glacial-marginal stream channels cut into bedrock on hillsides. If any ice lobes had filled the upper parts of these mountain valleys, such as Yellowstone Prong (Figures 1 and 8), traces of these channels should still exist. Commonly channels of this nature have a series of potholes developed along their bedrock floors and are found high above modern-day streams.

14. Till is one of the most characteristic deposits in glaciated areas. Regardless of the erosion and mass-wasting since late Wisconsin time, some patches of till should have been left preserved along the exposed courses of streams, in the lee of bedrock obstructions, in excavations for the Blue-Ridge Parkway or building sites, or exposed in other present day road cuts. Till has not been found in any of the above locations.

It would seem that snowbanks or snowfields seldom persisted throughout the summer season. Abundant block-stream or blanket-type boulder talus along many valley sides indicates a former more rigorous freeze-thaw environment during Wisconsin time. However, if glaciers had at one time occupied valleys at the highest elevations in this area, peaks above these valleys should have been modified into horns or pyramidal-shaped peaks. If glaciation was short-lived, horn-cutting would not have occurred. Such pyramidal-shaped peaks are missing from this region. If this area was covered by an ice-cap, which aggrevates this issue even more, then high peaks would be covered with glacial cobbles and boulders and exhibit frequent ice-molded forms. Figure 8 is a photograph of the east summit area of Black Balsam Knob (peak 6214 (1895 m) seen on Figure 1). Note the lack of a pyramid-shape of the upper summit cone and absence of cirque development on the favorable east side of this high ridge which would have been an optimum location for the accumulation of snow blowing across the flat upland which is hidden from view on this photograph. Also note the strong V-shaped cross-section profile of the valley extending toward the observer. This is the upper drainage basin of Yellowstone Prong (Figure 1) which in this view is above 5,400 feet (1646 m).

CONCLUSIONS

Several lines of evidence in the headwater reaches of the East Fork of the Pigeon River and Davidson River valleys demonstrate lack of mountain glaciation during late Wisconsin time.

a. There are no deposits of till.

b. There are no landscape features which resemble moraines even in valleys above 5,000 feet (1525 m).

c. Striations, grooves or other glacial markings can not be found across bedrock outcrops even at altitudes above 5,800 feet (1768 m) nor do striated, polished or faceted clasts occur in colluvium or stream deposits.

d. There is a notable absence of streamline molded forms.

e. Most ridge ends are not truncated.

f. Ice-contact gravel deposits do not exist even at high elevations.

g. Strong unequivocal cirque-forms are absent.

h. U-shaped cross-valley profiles are not found in any of the larger valleys.

i. Ice-marginal meltwater channels are missing.

j. Longitudinal profiles of major valleys lack eroded bedrock depressions.

k. Absence of high level gravel deposits suggest lack of glacial ice necessary to produce meltwater and detritus for deep valley fills.

l. The flanks of the highest mountains are not mantled by glacially transported boulders, patches of till, or ablation moraine.

m. Eskers, kames, or kame terraces are entirely absent.

n. Horn-shaped peaks do not exist.

o. Abundant talus suggests periglacial not glacial conditions.

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