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SOUTHEASTERN GEOLOGY

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LONG-TERM EPISODIC DEPOSITION ON MOUNTAIN FOOT SLOPES IN THE BLUE RIDGE PROVINCE OF NORTH CAROLINA: EVIDENCE FROM RELATIVE-AGE DATING

By

Hugh H. Mills
Department of Earth Sciences
Tennessee Technological University
Cookeville, Tennessee 38501

ABSTRACT

Bouldery diamictons on the piedmont slopes of higher mountains in the southern Blue Ridge province have been interpreted as the product of colluviation during Pleistocene glaciations. More recently, however, it has been demonstrated that similar sediments can be produced by debris flows set off by rare, catastrophic rainfalls, thus casting doubt on the climatic significance of foot-slope deposits. In the absence of any absolute dates, relative-age dating was used to shed light on the relationship of these deposits to climatic change. On the basis of weathering characteristics, soil-profile development, and topographic relationships, foot-slope deposits in the Dellwood, Grandfather Mountain, and Bakersville quadrangles of North Carolina were divided into several age groups. Then, based upon three weathering criteria (percent clay, Munsell hue, and clast weathering) measured in the B2 horizon at 135 sites, discriminant analyses were performed for each area. Plots of the scores on the resulting discriminant functions show discontinuities between the youngest and next older age group in all three areas. This quantitatively demonstrated gap in degree of weathering presumably reflects a time during which little or no deposition took place and thus indicates that deposition has been episodic on a time scale sufficiently great to allow substantial weathering to occur between successive episodes. This finding supports, although it does not prove, the concept that foot-slope deposition is related to Quaternary climatic change.

INTRODUCTION

Deposits of bouldery colluvium and alluvium are widespread on the piedmont slopes of higher mountains in the southern Blue Ridge province. These deposits often are referred to as alluvial or colluvial "fans." Many, however, lack a shape even crudely approximating that of a fan, and some occur as continuous belts along the bases of mountains. As the sedimentation processes probably are similar regardless of planimetric form, I shall refer to all such deposits collectively as *foot-slope* deposits. Their areas range from several hectares to several square kilometers, and well logs indicate that mean thicknesses rarely exceed 20 m. Commonly they are composed of a bouldery diamicton, although larger deposits usually display some fluvial gravels at their downstream ends.

Many observers have attributed these deposits to the effects of Pleistocene glacial climates. Michalek (1968), for example, thought that they were the product of gelifluction. However, numerous debris flows set off in central Virginia by catastrophic rainfalls during Hurricane Camille in 1969 demonstrated that many of the depositional features previously attributed to gelifluction can be produced by processes associated with the present climate. Hence, other observers (e.g., Gryta and Bartholomew, 1977, and personal communication, 1977) have contended that foot-slope deposits are continually being formed by floods and debris flows produced by rare catastrophic rainfalls, and thus are not necessarily related to climatic change. More recently, Kochel and others (1982) have obtained evidence for repeated Holocene debris avalanching in central Virginia, with radiocarbon dates indicating recurrence intervals of 700 to 4,000 yr. My own observations (Mills, 1981, and in press) indicate no convincing evidence for gelifluction. Even if most foot-slope deposits are the results of debris flows, however, this by no means eliminates the possibility that such deposition is controlled to some extent by climate. For example, debris flows may be larger or more frequent under certain climatic regimes than under others.

THE APPROACH

An obvious approach to the question of climatic control would be to see if the deposits date to times of glacial climates. The absence of absolute dates in the study areas, however, made this method impossible. The presence of foot-slope deposits of different ages may allow an alternative approach. In the Dellwood quadrangle, for example, Hadley and Goldsmith (1963) reported two ages of deposits and suggested that deposition occurred during early and late Wisconsinan. Note that if foot-slope deposits are climatically controlled in this manner, their ages should be grouped in time, being concentrated in intervals of favorable climate. If, however, the deposits are the products of catastrophic rainstorms randomly spaced through time, there should be no such grouping of ages. Although the existence of such grouping cannot be checked by absolute dating, relative-age dating may provide a means of doing so. This technique involves the quantitative measurement of weathering criteria believed to be a function of time; Colman and Pierce (1977), Burke and Birkeland (1979), and Miller (1979) list many weathering criteria that have been employed.

In order to see how this approach can be used in the present setting, consider the following hypothetical example. Suppose that foot-slope deposits at a large number of sites in a given area have been classified into one of two age groups, "older" or "younger," according to their weathering characteristics and topographic positions. The question arises of whether there are really two discrete age groups, or whether a continuum from least to most weathered has arbitrarily been divided into two parts. An answer can be obtained by making quantitative weathering measurements at each site and then plotting a frequency distribution of the resulting weathering index (Fig. 1). A result such as that in Figure 1A would suggest that no real grouping exists, whereas that in Figure 1B would suggest that grouping does exist. The latter result indicates episodic deposition with a time interval between successive episodes sufficiently long to allow substantial weathering to take place. Based on this rationale, an attempt was made to apply relative-age dating to a large number of foot-slope sites in the Blue Ridge province of North Carolina.

GEOLOGIC SETTING

Several areas described by Michalek (1968) as having well-developed "fans" were studied. These were the Dellwood quadrangle, the Grandfather Mountain area in the Grandfather Mountain quadrangle, and the area east of the Roan Mountain in the Bakersville quadrangle, all in North Carolina. The basic geologic features of each of these will be described briefly.

Dellwood quadrangle lies in the easternmost Great Smoky Mountains. According to Hadley and Goldsmith (1963), the bedrock is comprised of a thick sequence of metamorphosed sedimentary rocks of late Precambrian age, known as the Ocoee series, that rest on a "basement" complex of granite and metasedimentary gneisses. The Ocoee series consists of variously metamorphosed sandstone, siltstone, and mudstone. Generally rocks of the Ocoee series make up the mountain tops and the basement rocks the valleys and lower slopes, although in the Dellwood quadrangle this correlation is only approximate. The degree of metamorphism is high. The highest elevation in this area is about 1652 m, the maximum relief about 900 m, and the foot slopes are located at an elevation averaging about 1000 m.

The study area in the Grandfather Mountain quadrangle is entirely underlain by the mildly metamorphosed Grandfather Mountain Formation. According to Bryant and Reed (1970), arkose and subarkose make up about 50 percent of this formation, siltstone about 35 percent, and metamorphosed volcanic rocks the remaining 15 percent. Arkose (especially conglomeratic beds) generally underlies the peaks and siltstones the valleys and lower slopes. The highest elevation is about 1784 m, the maximum relief about 640 m, and the foot slopes occur at an elevation averaging about 1200 m.

No large-scale map has been made of the Bakersville quadrangle. However, the

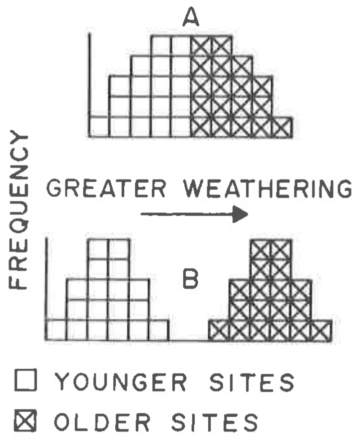

Figure 1. Histograms showing hypothetical frequency distributions of the weathering intensity measured at each of 36 sites according to some composite weathering index. Open blocks indicate younger sites, blocks with X's indicate older sites. Implications of the differences between distribution A and distribution B are discussed in the text.

Geologic Map of North Carolina (Stuckey and Conrad, 1958) shows the western slopes of Roan Mountain to be underlain predominantly by hornblende gneiss and granite gneiss, and to a lesser extent by gabbro. Generally hornblende gneiss underlies the upper slopes, granite gneiss the lower. The highest elevation is 1911 m, the maximum relief greater than 1200 m, and the foot slopes occur at an elevation averaging about 950 m.

PROCEDURE

Between 6 and 13 foot slopes were visited within each study area, several sampling sites being located on most. No substantial systematic variation in slope, climate, parent material, or other factor likely to bias the results was apparent in any area. There were 28 sites in the Dellwood quadrangle, 27 at Grandfather Mountain, and 80 at Roan Mountain. Sites were located at road cuts or other excavations that gave access to the subsoil. Within each area, sites were initially classified into relative-age categories on a semi-objective basis, using weathering characteristics and topographic positions. Because of the relatively small number of sites, only two age categories were used at Dellwood and Grandfather Mountain, whereas three were used at Roan Mountain. There appeared to be fairly clear breaks between the two groups at Dellwood and Grandfather Mountain. At Roan Mountain the break between young and intermediate ages was also fairly clear, but that between intermediate and older seemed more arbitrary.

The next task was to determine whether the age groupings and the apparent breaks between them could be demonstrated in a quantitative manner. Three measures of weathering intensity were used: the percent clay in the less-than-2-mm fraction, the extent of oxidation reflected by the reddest hue, and clast weathering, all measured in the B2 horizon. The hue (Munsell, 1975; moist condition) was measured in the lab under natural lighting, using a randomized order of analysis. Clast weathering was measured in two ways. At Roan Mountain the abundant hornblende-gneiss and amphibolite clasts displayed relatively well-defined weathering rinds, although somewhat more gradational than those of basalt and andesite used in previous studies (e.g., Colman and Pierce, 1980). A minimum of 10 clasts at each site were randomly selected and the mean thickness of the weathering rind was determined. At Dellwood and Grandfather Mountain the metasedimentary clasts showed poorly defined rinds, so that a different measure of weathering was used. A total of 25 pebble-size clasts were randomly selected and each classified as to whether it could or could not be broken apart by hand (into pieces granule size or smaller), and the "percent weathered clasts" then recorded. Using these three variables I then tested my relative-age classification in each area by means of three discriminant analyses (Klecka, 1975).

RESULTS

The mean values of the weathering variables (Table 1) show large differences between age groups. From what is known of the rate at which weathering occurs (e.g., Birkeland, 1974, p. 153-180), differences of this magnitude probably require a minimum of 10,000 yr to develop. The question remains, however, of whether the breaks between the groups are real. In Figure 2, the results of the discriminant analyses are presented as frequency histograms of the discriminant-function scores. The score on the discriminant function is analogous to the hypothetical weathering index in Figure 1. (For mathematical reasons of no importance for interpretation, greater weathering is indicated on Fig. 2 by a more negative score.) The histograms in Figure 2 obviously are more complex than the hypothetical ones in Figure 1, as they show numerous gaps. However, many of these gaps are defined by only two or three sites and probably have little meaning. The proper way to view the histograms is to confine attention to the broad groupings, seeing whether the largest breaks occur between the previously defined relative-age groups. On the Grandfather Mountain histogram, for example, there are obviously fewer sites having discriminant-function scores between 1.6 and -1.0 than there are above and below this range. Similarly, on the Dellwood histogram, although the older sites show a wide range, the gap between 0.00 and -1.4 indicates that these sites are distinct from the younger sites. On the Roan Mountain histogram, there is a break between about 0.2 and 1.0 separating the young and intermediate sites. A break between the intermediate and old sites is questionable, however.

Thus, the histograms show that in all three areas a discontinuity exists between the youngest sites and the next older ones. Admittedly the breaks are not as dramatic as one might wish, and if only one area had been studied the existence of a break might be disputable. The fact that all three areas show such breaks, however, makes the results difficult to attribute to coincidence alone.

Table 1. Means and standard deviations of weathering measurements on materials from the B horizon.

Area	Age	N	Percent Clay	Hue (YR)	Percent Weathered Clasts	Mean Rind Thickness (mm)
Dellwood	Young	7	23.0 ± 9.3	9.76* ± 0.75	8.2 ± 9.8	
	Old	21	32.0 ± 9.4	5.71 ± 2.78	60.0 ± 14.4	
Grandfather Mountain	Young	11	28.3 ± 6.9	10.00 ± 0.00	4.0 ± 5.1	
	Old	16	26.0 ± 8.6	5.78 ± 2.54	58.3 ± 24.1	
Roan Mountain	Young	26	22.9 ± 11.8	8.22 ± 1.07		0.29 ± 0.21
	Intermed.	26	35.4 ± 6.7	6.39 ± 1.29		1.40 ± 0.70
	Old	28	50.3 ± 11.4	4.95 ± 1.58		5.04 ± 2.24

*All hue values were YR (Munsell). Therefore, as hue is a continuous variable measured on an interval scale, the hue values could be treated the same as values of any other continuous variable.

DISCUSSION

Although the histograms in Figure 2 suggest that at least two discrete age groups of deposits are represented, one may ask why the discontinuities are not clearer. An important reason probably is that the relative-age criteria employed are not perfect indicators of age, likely sources of error being small lithologic and microenvironmental differences between sites. Perhaps such differences become more important with greater age, thereby obscuring the relative-age grouping of older deposits. Another reason is that although depositional events are concentrated in certain time intervals, probably they are by no means confined to these intervals, with some deposition occurring at other times.

Demonstration of long-term intermittent deposition does not, of course, necessarily prove climatic control. Intermittent uplift is a possible cause, although this does

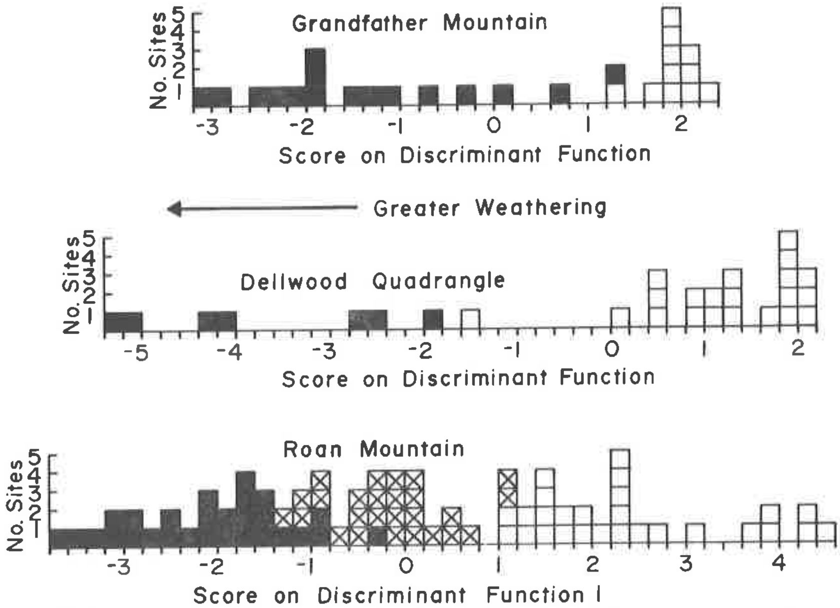

Figure 2. Histograms showing scores on discriminant function for each site in each of the three study areas. Open blocks represent young sites and filled blocks, old sites. At Roan Mountain the blocks with X's indicate sites of intermediate age. Note that on this figure weathering increases to the left. For the Roan Mountain sites, percent clay and mean rind thickness values were transformed logarithmically before the discriminant analysis was performed.

Note that where there are only two groups (Dellwood and Grandfather Mountain) there is only one discriminant function, and the resulting plot is a histogram, as shown. Where there are three groups, however, two discriminant functions are derived, and normally the results are shown by plotting the score on the first function against the score on the second. In the case of Roan Mountain, however, a very low eigenvalue indicated that the second function has almost no discriminative power. Thus, scores on the first function have been plotted in histogram form in the same manner as for the other two study areas.

not seem very likely in an area of generally low tectonic activity such as the Appalachians. Another possibility is the occurrence of catastrophic rainstorms with recurrence intervals on the order of 10,000 yr, rather than the much shorter intervals suggested by Kochel and others (1982), as intervals of this magnitude would be necessary to allow the observed weathering differences between age groups to develop. This interpretation, however, would require deposits of a given age to be the product of one depositional event, and the deposits seem much too voluminous to be accounted for in this manner. Williams and Guy (1973), for example, reported that the average thickness of foot-slope deposition produced by the Hurricane Camille floods was only 30 to 60 cm. It seems more reasonable to attribute deposits of a given relative age to a large number of depositional events relatively closely spaced in time (say, within an interval of several thousand years).

If climatic change is accepted as the cause of the intermittent foot-slope deposition, the mechanisms and timing of deposition still must be determined. One commonly held assumption is that deposition takes place during glaciations and incision during interglaciations. However, Delcourt's (1980) radiocarbon-dated sequences of terrace alluvium along the Little Tennessee River in east Tennessee suggest that episodes of massive alluviation (and by implication, colluviation) in the southern Appalachians are confined primarily to lateglacial/interglacial (and stadial/interstadial) transitions. Thus, the younger group of sites presumably is lateglacial to early Holocene in age. The older deposits, exhibiting a greater range of weathering intensity, probably correlate with pulses of sedimentation occurring as recently as mid-

Wisconsinan, although many probably are associated with somewhat older episodes. Undoubtedly some deposits date from other time intervals, including late Holocene.

CONCLUSIONS

The results suggest that foot-slope deposition has been intermittent on a time scale sufficiently great to allow substantial weathering to take place between depositional episodes. The most reasonable explanation of such intermittent behavior is the effect of Quaternary climatic cycles. However, proof of this inference and precise timing of the events await absolute dating from sites throughout the region.

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STRUCTURE AND GEOLOGIC HISTORY OF A PART OF THE CHARLOTTE BELT, SOUTH CAROLINA PIEDMONT

By

Neil J. Gilbert
Law Engineering Testing Company
P. O. Box 11297
Charlotte, North Carolina 28220

Henry S. Brown
Department of Marine, Earth and Atmospheric Sciences
North Carolina State University
Raleigh, North Carolina 27650

Malcolm F. Schaeffer
Duke Power Company
P.O. Box 33189
Charlotte, North Carolina 28242

ABSTRACT

Several fault zones were discovered during geologic mapping of a construction excavation located in the Charlotte belt of northeastern York County, South Carolina. The faults and other structural elements were studied to develop the geologic history of the site and the surrounding area. The investigation included detailed geologic mapping, almost 2500 measurements of structural elements, study of more than 200 thin sections and radiometric dating of mineral separates.

Adamellite crystallized penecontemporaneously with formation of mafic, aplite and pegmatite dikes. Deformation, including faulting, accompanied crystallization and later metamorphism to the amphibolite facies. Subsequent greenschist facies metamorphism was followed by renewed faulting. A sequence of hydrothermal mineralization followed the last faulting. Prehnite-calcite and laumontite-calcite veinlets parallel and cut across shear fractures of the fault zones. An essentially pure separate of laumontite was dated at 86 ± 30 million years by the potassium-argon method.

The angular relationships between faults approximate a theoretical pattern for a wrench fault system based on the existence of brittle conditions. However, the parallelism of the largest faults to mafic dikes and the branching, anastomosing nature of the shears attest to an original condition of faulting that was not purely brittle. Later deformation was controlled as much by the orientations of existing planes of weakness as by the prevailing stress field.

INTRODUCTION

Structural and petrologic information in the Charlotte belt of the Piedmont is difficult to obtain because of generally poor exposures. Recently completed bedrock mapping of a large excavation has provided a unique opportunity to examine the petrology, structure and geologic history of a part of the Charlotte belt in the South Carolina Piedmont. The results of such a study are of fundamental importance to a better understanding of the regional geology.

Several faults were discovered during geologic observation and mapping of a large construction excavation located in northeastern York County, South Carolina (Figure 1). Geologic investigation of the site between mid-1975 and mid-1978 included detailed mapping of areas totalling more than 83,500 sq m (900,000 sq ft) at scales of 1 in equals 1, 5 and 10 ft, almost 2500 Brunton compass measurements of structural elements, examination of more than 200 thin sections of rock and saprolite, X-ray diffraction analyses of mineral specimens and radiometric dating of selected materials by the uranium-lead and potassium-argon methods.

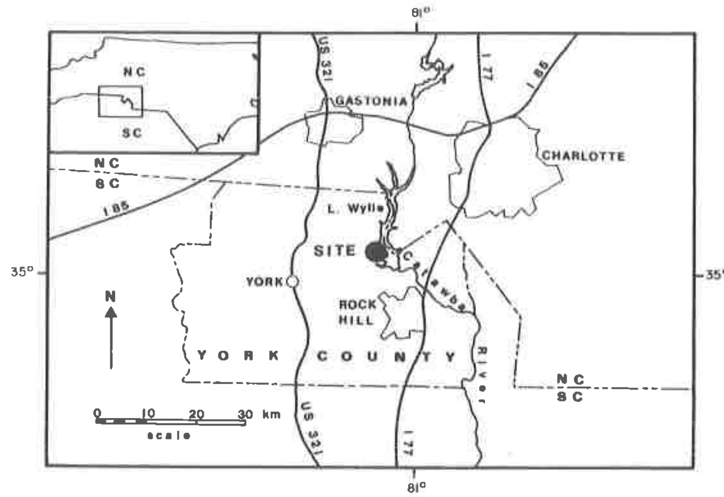

Figure 1. Location map.

GEOLOGIC SETTING

The Piedmont Physiographic Province is a deeply eroded, plateau-like segment of the Appalachian Mountain System and in this region is about 130 to 190 km wide. The Charlotte belt (Figure 2), one of four geologic belts within the Piedmont, extends from near Winston-Salem, North Carolina into South Carolina (King, 1955). It has a width of 50 to 65 km. Contained within this belt is a complex series of intrusive rocks, with some schist, quartzite, gneiss and amphibolite probably derived from sedimentary and volcanic deposits (Butler, 1966). Metamorphic rocks are mainly amphibolite facies. Intrusive rocks range in composition from granite to gabbro; some of the granitic bodies are of batholithic dimensions (King, 1955; Butler and Ragland, 1969). It is mainly the extensive complex of intrusive rocks which distinguishes the Charlotte belt from the adjacent geologic belts.

PLUTONIC AND METAMORPHIC HISTORY

Plutonism

Three principal episodes of plutonism are delineated in the Charlotte and Carolina slate belts (Butler and Ragland, 1969; Fullagar, 1971; Fullagar and Butler, 1979): (1) Pre-metamorphic plutons range in composition from granite to gabbro and have ages of 595 to 520 million years (Precambrian-Cambrian), (2) Granite plutons, some probably syn-metamorphic and some post-metamorphic, and a group of post-metamorphic gabbro-diorite-syenite plutons have ages of 415 to 385 million years (Silurian-Devonian), and (3) Post-metamorphic granitic intrusions are about 300 million years old (Pennsylvanian). North- to northwest-trending diabase dikes of Mesozoic age are widespread in the region.

The oldest and most abundant rock in the site area is metamorphosed adamellite. [The authors recognize that several igneous compositions are included within the broad term "adamellite" used for this mapping. Most of the rock so designated would classify as granite, granodiorite and quartz monzonite under the new IUGS system (A.L. Streckeisen, "Plutonic Rocks - Classification and Nomenclature Recommended by the IUGS Subcommission on the Systematics of Igneous Rocks," Geotimes, p. 26-30, October, 1973). The single term "adamellite" was retained because of the relative uniformity of the rock in hand specimen and for consistency with earlier published and unpublished field maps.] The rock ranges from well foliated to massive in appearance; there is normally some evidence of foliation. Adamellite found at many localities in York County is cut by mafic dikes that have been metamorphosed to the amphibolite facies (Butler, 1966; written communication, 1975). The strike of these dikes is

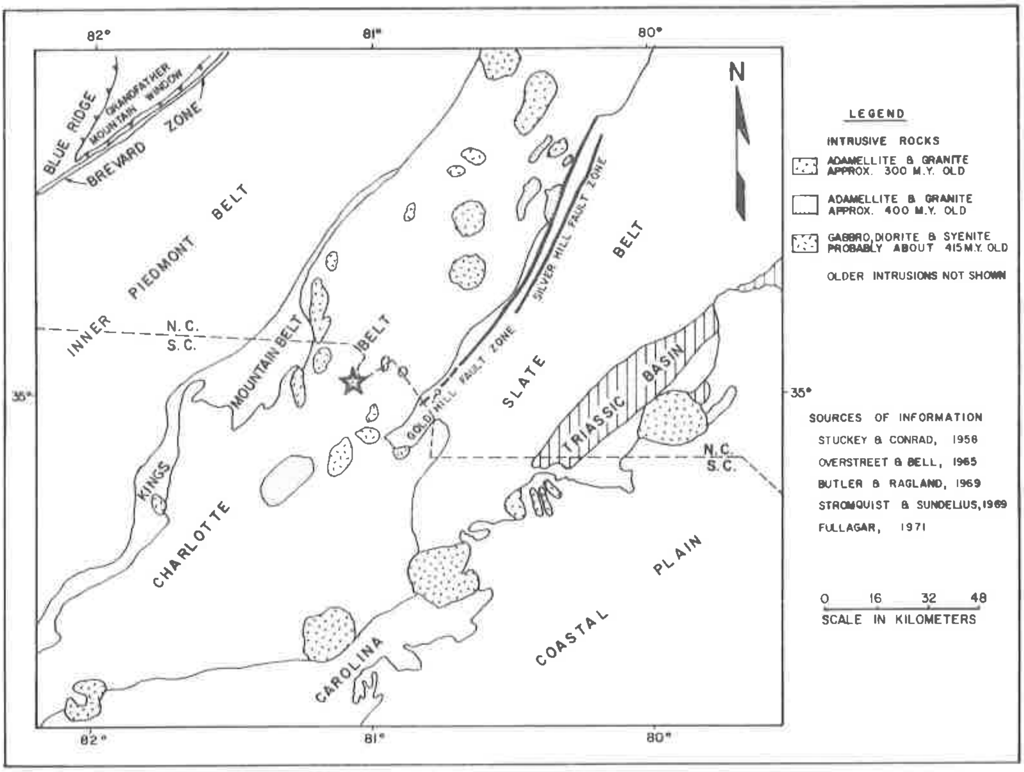

Figure 2. Geologic setting. Location of site indicated by star.

commonly north to northeast.

The predominant rock type underlying the site is adamellite. The earliest post-adamellite rocks are aplite and pegmatite dikes. Aplite dikes range in length from 2 to 40 m (average about 10 m) and are up to 4 m thick (average about 10 to 15 cm); pegmatites range from 1/2 to 230 m long (average about 25 m) and up to about 4 m thick (average less than 1/2 m). These early dikes are cut by and often offset along the slightly later mafic dikes (Figure 3). Mafic dikes range in length from 2 to 120 m (average about 40 m) and in thickness up to 4 m (average about 1 m). At some locations observed, mafic dikes have a round, bulbous form. Based on field observations, mafic dikes apparently were separated within a host material that was still plastic or in a crystal mush condition with adamellite flowing into the spaces created by the separation. Thus, most rocks at this site appear to have crystallized penecontemporaneously. The age of crystallization of the adamellite is 532±15 million years based on a uranium-lead determination of zircon (Table 1).

Amphibolite-Facies Metamorphism

Regional metamorphism reached its peak 420 to 380 million years ago (Fullagar, 1971; Butler, 1972) and imparted a moderate foliation to the mafic dikes and a faint foliation to the adamellite. The foliation is older than the nearby Lowrys pluton (29 km southwest of the site) dated at 399±4 million years (Butler, 1971; Fullagar, 1971, 1981). Subsequent movement deformed the foliation in the mafic dikes at their contacts. Stretching and necking of some mafic dikes occurred and pegmatite stringers developed at the mafic dike pinchouts.

Mineral assemblages of amphibolite facies are present in both major rock types at the site. Mafic dikes consist essentially of fine-grained plagioclase, hornblende and biotite. Plagioclase, making up almost one-half the rock, is andesine-labradorite. Hornblende is dark green in stubby, euhedral to subhedral grains. Biotite, making up no more than 10 to 15 percent of the rock, occurs as subhedral to euhedral grains in

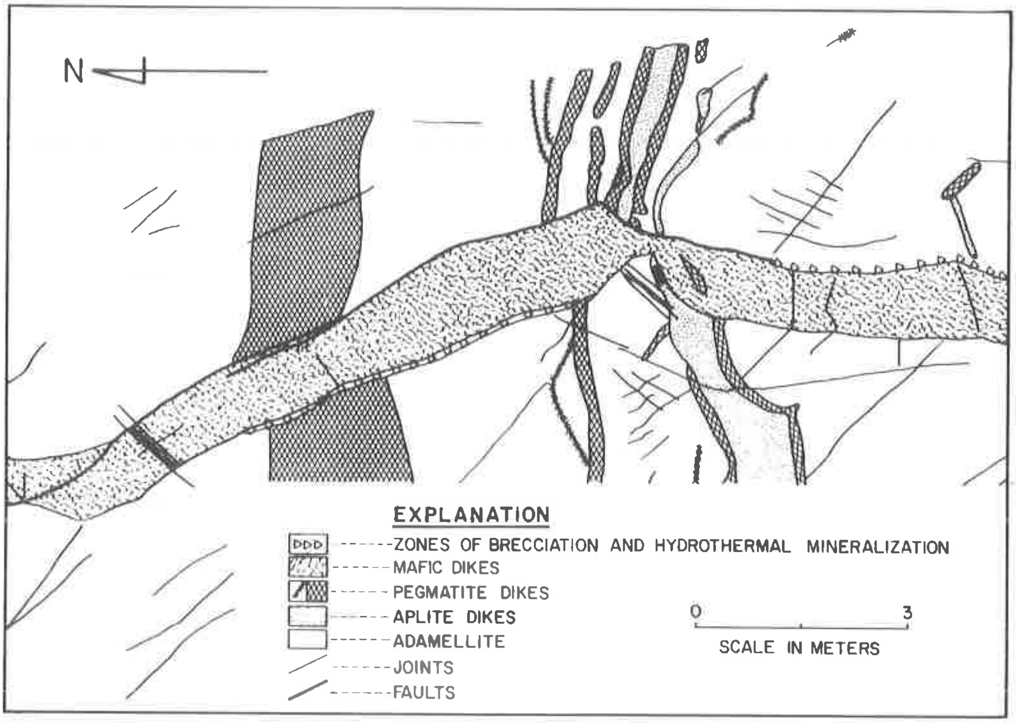

Figure 3. Geologic map showing aplite and pegmatite dikes offset along mafic dike.

Table 1. Radiometric Age Determinations.

Uranium-Lead Age of Zircon in Adamellite

	Measured Atomic Ratios			Corrected Ratios			Calculated Ages (Millions of Years)		
Sample Designation	$^{208}Pb/^{206}Pb$	$^{207}Pb/^{206}Pb$	$^{204}Pb/^{206}Pb$	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$	$^{207}Pb/^{206}Pb$	$^{206}Pb/^{238}Pb$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
Boring Z3-B1 101.2 - 105.2 ft.	0.1972	0.06235	0.000340	0.5376	0.06795	0.05740	424	437	510
Boring Z3-B1 101.2 - 105.2 ft. (Leached in HF)	0.1626	0.06150	0.000238	0.6560	0.08203	0.05803	508	512	530

Ratios Used For Common Pb: $^{208}Pb/^{204}Pb = 38.20$, $^{207}Pb/^{204}Pb = 15.72$, $^{206}Pb/^{204}Pb = 18.51$

Decay Constants Used: $\lambda\ ^{238}U = 0.1551 \times 10^{-9}\ yr^{-1}$, $\lambda\ ^{235}U = 0.9848 \times 10^{-9}\ yr^{-1}$

Crystallization (Concordia) Age 532 ± 15 Million Years

Potassium-Argon

Sample Designation	Material Analyzed		Potassium (Weight Percent)	$^{40}Ar \times 10^{-4}$ Scc/gm	Percent $^{40}Ar_{rad}$	Calculated Age (Million of Years)
Boring Z3-B1 89.4 to 91.0 ft.	Laumontite		0.65	0.02302	53.0	86 ± 30
Boring A-72 55.0 to 55.5 ft.	Amphibolite	Biotite	5.69	0.738	97.3	300 ± 5
		Hornblende	0.82	0.822	82.2	316 ± 9
Boring A-32 88.0 to 88.4 ft.	Amphibolite	Hornblende	0.74	0.1001	85.0	310 ± 12
		Whole Rock	1.23	0.160	93.3	301 ± 7
Boring A-32 125.8 to 126.1 ft.	Adamellite	Biotite	7.64	1.069	95.4	322 ± 8
		Whole Rock	1.36	0.2139	90.8	359 ± 11

Constants: $^{40}K/K = 1.187 \times 10^{-4}$ mol/mol

$\lambda_{\beta} = 4.72 \times 10^{-10}\ yr^{-1}$

$\lambda_{\epsilon} = 0.584 \times 10^{-10}\ yr^{-1}$

NOTE: All Radiometric Analyses Performed by A. L. Odom (Florida State University).

subparallel alignment imparting a weak foliation to the rock. Adamellite is medium grained and contains approximately 40 percent plagioclase (oligoclase), 20 percent potash feldspar, 20 percent quartz, 10 percent biotite and 5 percent hornblende. Epidote, sphene, apatite and some iron oxides occur as common accessories in both rock types. Hornblende may be igneous. Some or all biotite is metamorphic as suggested by its observed replacement of hornblende and its preferred orientation.

The equilibrium mineral assemblage in these rocks, hornblende-medium plagioclase-epidote-biotite ($\pm$quartz), is characteristic of the amphibolite facies (Turner, 1968). The minimum temperature of the amphibolite facies is in the neighborhood of 550^{o} C and is only slightly pressure dependent (Miyashiro, 1973). The coexistence of medium plagioclase (oligoclase-labradorite) with epidote, however, suggests relatively high pressure (Turner and Verhoogen, 1960). Thus, ranges of 550^{o} to 650^{o} C at 400 to 700 MPa (58,000 to 101,000 psi) in the site area probably are the temperature and pressure conditions that resulted in metamorphism to the amphibolite facies.

Greenschist-Facies Metamorphism

Minor amounts of later-formed minerals occur mainly in and near shear and breccia zones. Quartz-epidote veinlets cut across foliation of metamorphosed adamellite. Original minerals in adamellite are replaced to varying degrees by muscovite, chlorite, epidote and calcite, and portions of amphibolite dikes are altered to chlorite-rich and epidote-rich assemblages. This assemblage, calcite-epidote-muscovite-chlorite-quartz, is typical of the greenschist facies (Turner, 1968). Albite has not been identified, but very fine-grained, sugary-textured minerals occurring in and possibly replacing portions of shear zones may be mixtures of quartz and albite. Plagioclase is commonly altered to fine-grained muscovite while microcline is relatively free of alteration effects. All colorless micas are herein called muscovite even though some may be other types, such as phengite, that are indistinguishable from muscovite in thin sections. The alteration of plagioclase and hornblende releases calcium, some of which may be fixed as calcite, and probably accounts for calcite masses and veins especially common in and near shear and breccia zones. Olive-green biotite is altered to reddish brown biotite with release of iron oxide. Chlorite and fine-grained iron oxide replace biotite and wallrock adjacent to fractures. Iron oxide, identified by X-ray diffraction as mainly hematite, is a major component of some of the small shear zones and is associated with greenschist mineralization. The temperature under which quartz-epidote veinlets, breccia zones and iron oxide-defined shear zones and associated mineralization occurred was in the range of 350^{o} to 550^{o}C at a pressure of 300 to 600 MPa (43,000 to 86,000 psi) (Miyashiro, 1973).

There are two possible interpretations for the onset of greenschist metamorphic conditions: (1) Fracturing late in the main metamorphic event permitted invasion of water-rich fluids and formation of hydrous, lower grade minerals, or (2) Retrogression accompanied fracturing in a separate metamorphic episode distinctly later than the main regional event. The general limitation of greenschist mineralization to fractures and shear-breccia zones suggests that greenschist metamorphism is related to fracturing. Work by others in the Piedmont (Fullagar, 1971; Butler, 1972; Bobyarchick and Glover, 1979) has resulted in similar interpretations. Thus, there is mounting evidence that greenschist mineralization of the type described here, though it may be hydrothermal in nature, is regional in extent.

Potassium-argon dates on separated minerals and whole-rock samples range from 359$\pm$11 to 300$\pm$5 million years (Table 1). The results of isotopic analyses, when plotted on an isochron diagram (Figure 4), yield mineral-whole rock isochrons of about 300 million years. The three different types of materials analyzed (biotite, hornblende and whole-rock samples) have significantly different closing temperatures to migration of argon. The fact that these three materials define isochrons of 300 million years indicates a rapid cooling rate. This may suggest a more rapid uplift than previously reported in the area (Butler, 1972) or the ages may represent rapid cooling after the intrusion of granitic plutons about 300 million years ago (Fullagar, 1971).

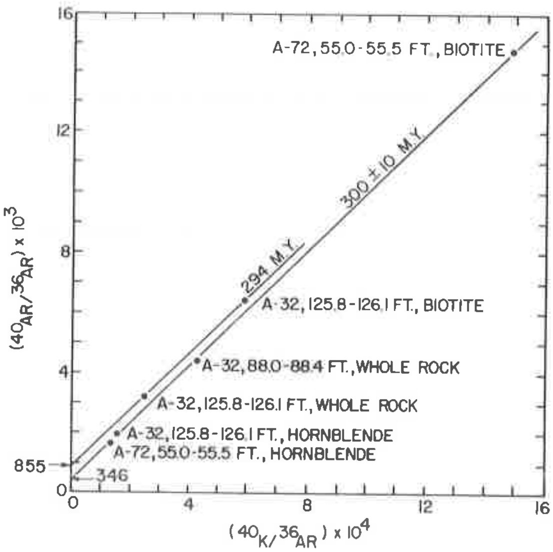

Figure 4. Isochron diagram. Results of potassium-argon age determinations of whole-rock samples and mineral separates.

Hydrothermal Mineralization

The youngest mineral assemblages include prehnite, laumontite and calcite; they occur in and near dilation fractures. Similar assemblages introduced into fractures by hydrothermal solutions have been studied by Thompson (1971). Field and experimental data indicate that prehnite-bearing assemblages generally form under conditions of higher temperature and pressure than laumontite-bearing assemblages (Liou, 1971; Zen and Thompson, 1974). At this site, laumontite-calcite veins occasionally cut across slightly earlier prehnite (Figure 5). Some textures in thin section suggest that laumontite has partially replaced prehnite. Winkler (1976) has proposed the equilibrium reaction, prehnite + quartz + $3H_2O$ + CO_2 = laumontite + calcite; this could account for the observed relationship assuming a slight increase in P_{CO_2} with time. Temperature and pressure conditions under which laumontite would be expected to form are in the range of 160° to 300° C and 100 to 300 MPa (14,000 to 43,000 psi) water pressure (Thompson, 1971; Zen and Thompson, 1974).

A date of 86±30 million years was obtained from a potassium-argon analysis of laumontite from filled fractures in rock core (Table 1). The sample analyzed was hand-picked under a binocular microscope so that each grain was examined. Grains which contained fragments of other minerals were rejected. Sample purity was confirmed by X-ray diffraction analysis of both moist and dry splits (A.L. Odom, written communication, 1976). The laumontite sample has 0.65 weight percent of potassium, which is within the range of 0.30 to 0.66 percent potassium reported by Deer, Howie and Zussman (1969). They noted that "the majority of reliable analyses of laumontite show an appreciable content of alkalies" (p. 402). Thus, the laumontite dated for this investigation does not have an anomalous potassium content. Laumontite is a tectosilicate with a relatively open lattice structure and is subject to loss of argon. Therefore, we interpret this analysis to be a minimum age.

Occurrences of laumontite and related minerals (prehnite is a common associate) are widespread in the Piedmont (Privett, 1974). Laumontite generally occurs as fracture fillings and in hydrothermally altered zones adjacent to the fractures. Near Durham, North Carolina, laumontite was found in a fracture in Mesozoic diabase (Furbish, 1965) and near Leesburg, Virginia, prehnite and zeolites occur in fractures and vesicles in Mesozoic diabase (Toewe, 1966). The association of zeolites and related minerals with diabase suggests that the hydrothermal mineralization is concentrated in the early-to-mid-Mesozoic (180 to 220 million years) fractures along which the diabase was introduced (Butler, 1977). Therefore, the zeolites associated with Mesozoic (150

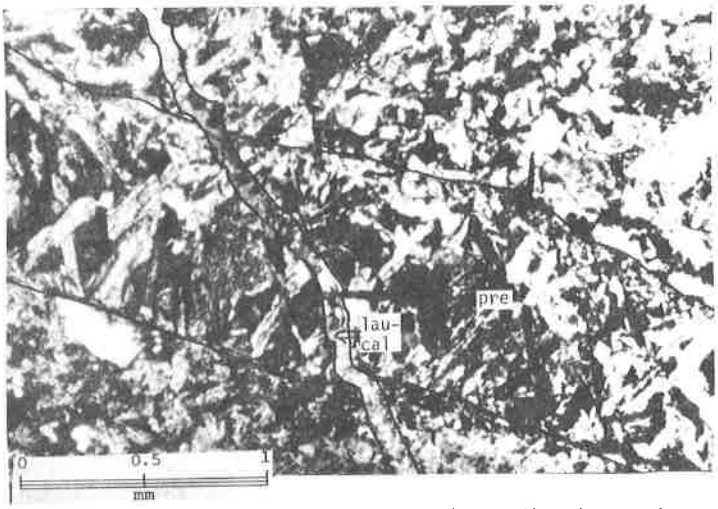

Figure 5. Photomicrograph of laumontite-calcite (lau-cal) vein cutting prehnite (pre) vein in adamellite.

to 220 million years) fracturing and intrusion are probably of equivalent age and not less than about 150 million years. In the absence of evidence to the contrary, it is speculated that other occurrences of zeolites in the Piedmont have similar origin and age (see, for example, Butler, 1977; Bobyarchick and Glover, 1979). The best interpretation of the potassium-argon age obtained during this investigation is that there has been selective loss of radiogenic argon relative to potassium and that the 86±30-million-year age is a minimum age for formation of the laumontite; its true age is probably not less than about 150 million years.

HISTORY OF FAULTING AND FRACTURING

Late-Magmatic Deformation

Evidence of the earliest faulting at the site consists of pegmatite dikes that are offset where they are cut by mafic dikes (Figure 3). In some cases, there are no fault textures where these offsets occur, suggesting that flow rather than fracturing was involved. In other places, faulting is recorded as stretched fragments along some of the mafic dike contacts. Such faults tend to disappear along strike into undeformed rock. It is interpreted that this deformational event occurred late in the magmatic episode, during crystallization of the adamellite and emplacement of the aplite, pegmatite and mafic dikes.

Deformation continued after the rocks had crystallized and resulted in textures characteristic of two types of fault processes: shearing and brecciation. Ductile shearing appears to be the earlier mechanism of faulting; some breccia zones cut across shear zones and some contain randomly oriented fragments of sheared rock. Shearing may have been initiated as one of the last effects of plutonic activity or as deformation that accompanied metamorphism to the amphibolite facies.
Brecciation, or brittle deformation, occurred during late phases of amphibolite metamorphism and during greenschist metamorphism. These two fault textures are discussed in the following paragraphs.

Shearing

In exposures of rock, shear zones range in thickness from a fraction of a centimeter to about one meter. The thicker shear zones contain many shear planes, generally concentrated near the centers of zones. Individual shear planes cut through and offset pegmatites and mafic bodies. Hematite (and other iron oxides) commonly occurs in very thin bands within and parallel to shear "planes."

In thin section, some ductile shear zones have mylonitic textures (Figure 6) and others are marked by anastomosing shear planes. These planes are not continuous, but curve, branch and disappear. Rock adjacent to shear zones is commonly strained or slightly brecciated. Shear zones are often marked by ribboned quartz and by iron oxide films outlining individual shear planes. Iron oxide films, probably introduced into the planes during greenschist alteration, may be closely spaced making a rather opaque zone or may alternate with rock slices locally imparting an augen texture.

Brecciation

Breccia zones occur in adamellite and along the contacts of some mafic dikes. They consist of angular to subangular fragments enclosed in a fine-grained or granular matrix. Bent muscovite crystals and round and elongate epidote crystals occur in the breccia. Quartz has recovered from strain through polygonization. Small, euhedral epidote crystals are intergrown with larger deformed ones in the breccia matrix and indicate the end of brecciation and greenschist metamorphism.

Faulting occurred along low-angle planes after brecciation. Where offset could be determined, net slip appears to be reverse-oblique up to 75 cm along planes usually dipping from 4 to 10 degrees (strike varies widely). These low-angle faults offset earlier shears and fractures filled with greenschist facies minerals, but are themselves truncated by faults and fractures filled with zeolites and/or other hydrothermal minerals.

Reactivations of older (pre- or syn-metamorphic) fault zones followed; quartz-epidote veinlets are offset up to 90 cm along some zones. In thin section, this late faulting is seen as thin seams of fine-grained, brecciated materials within greenschist-mineral-healed fault zones.

Dilation

The final deformation was the development of dilation fractures during an episode of hydrothermal activity. Some or all of the dilation fractures may represent reopening of earlier joints and shear planes. Prehnite and calcite were the first hydrothermal minerals to be introduced into these fractures. Prehnite crystals formed on fracture walls and grew inward, often leaving voids in vein centers. Calcite commonly accompanies prehnite and is seen in thin sections of fresh rock but not in thin sections of weathered rock or saprolite. Dilation fractures containing prehnite and calcite cut

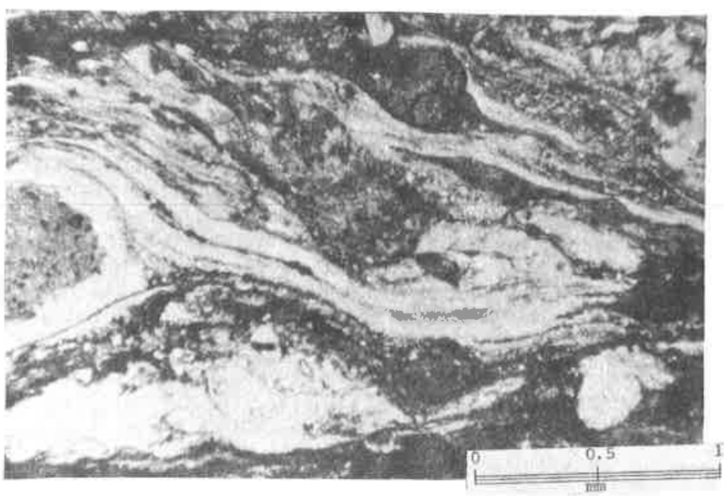

Figure 6. Photomicrograph of mylonite developed in adamellite. Augen of epidote (dark) and sericitized feldspar (gray) are surrounded by ribboned quartz.

across all shear and breccia zones that underwent post-greenschist deformation (Figure 7). Prehnite replaced minerals in rock within a few millimeters of prehnite-calcite veins and also formed within parts of fine-grained breccia matrix. In these situations prehnite and calcite cut across and replace greenschist minerals.

Following prehnite-calcite emplacement, laumontite and calcite were formed. In some cases laumontite and calcite filled open spaces in prehnite veins. In other cases they filled dilation fractures that cut across all older features including shear zones, breccia zones and prehnite-calcite veins (Figure 8). Dilation must have continued during zeolite crystallization, for some fractures filled with slightly strained laumontite have been expanded by dilation fractures that were filled with later, undeformed laumontite. Following laumontite-calcite mineralization, trace amounts of lower temperature-pressure zeolites such as stilbite and chabazite formed in open fractures.

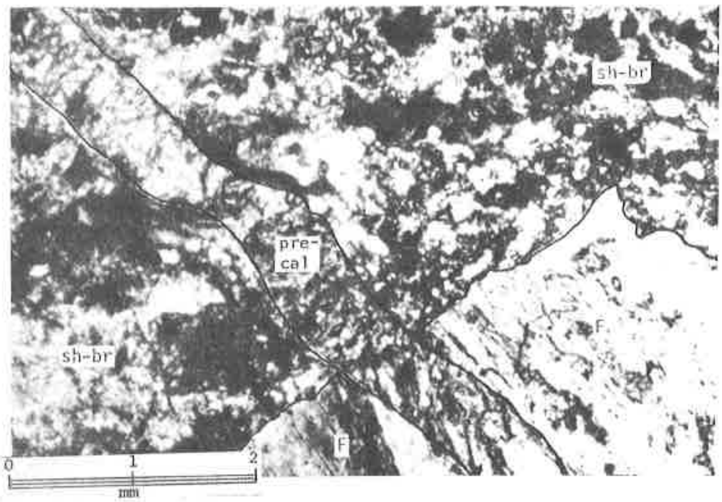

Figure 7. Photomicrograph of prehnite-calcite vein cutting shear-breccia zone. Shear-breccia zone (sh-br); feldspar (F); prehnite-calcite (pre-cal).

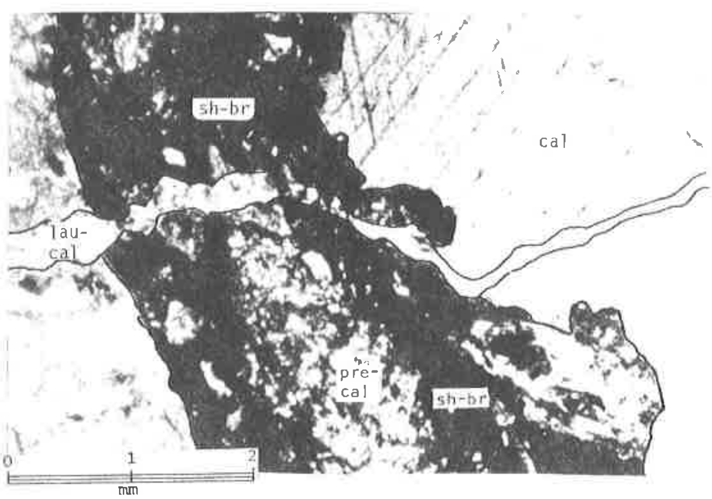

Figure 8. Photomicrograph of laumontite-calcite (lau-cal) vein cutting all earlier features including shear-breccia (sh-br) zone, calcite (cal) and prehnite-calcite (pre-cal) vein.

STRUCTURAL ANALYSIS

General observations of the pattern of faulting in the excavation include:

1. Shear fractures occur predominantly in four orientations: north-northeast, north, north-northwest and northwest.
2. The north- and north-northeast-trending sets are distinctly longer than the north-northwest and northwest-trending sets.
3. The north-, north-northwest- and northwest-trending sets each contain more numerous individual shear planes than the north-northeast-trending set.
4. About 70 percent of the northwest-trending shear fractures have right-lateral offsets. More than 80 percent of the remaining shears are left-lateral.
5. Apparent horizontal offsets range from a fraction of a centimeter to more than 6 m. The largest occur on the north and north-northeast trends.

These general observations are discussed in the following paragraphs.

Structural Framework

Structural measurements (approximately 2500 total) were made on shear planes, joints, foliation, and contacts of mafic, pegmatite, and aplite dikes during detailed geologic mapping of the site. The primary orientations of shear planes are north and northwest (Figure 9A). Primary joint orientations are northeast and northwest (Figure 9B). Mafic dike contacts are principally oriented north to north-northeast (Figure 9C). Dips of all planar features are generally greater than 70 degrees. There is a general parallelism among these structural elements in three orientations; northwest, north and north-northeast.

The largest faults observed, as defined by total length and amount of offset, occur along contacts of mafic dikes. Shearing occurred along highly irregular surfaces on one or both sides of mafic dikes. Where net slip could be calculated, it ranged from about 2 to 7 m, with the strike-slip component about twice that of the dip-slip. Strike-slip offset generally varies from millimeters to several centimeters, with the largest having an apparent left lateral offset of 6 m. The outcrop pattern of shears in the excavation is shown on Figure 10 and a straight-line generalization is shown on Figure 11. The four predominant shear plane orientations can be seen on these figures and in general, the north- and north-northeast-trending shears are distinctly longer than the northwest- and the north-northwest-trending shears.

Structural Synthesis

The fault system was studied by referencing both orientation and length of shear

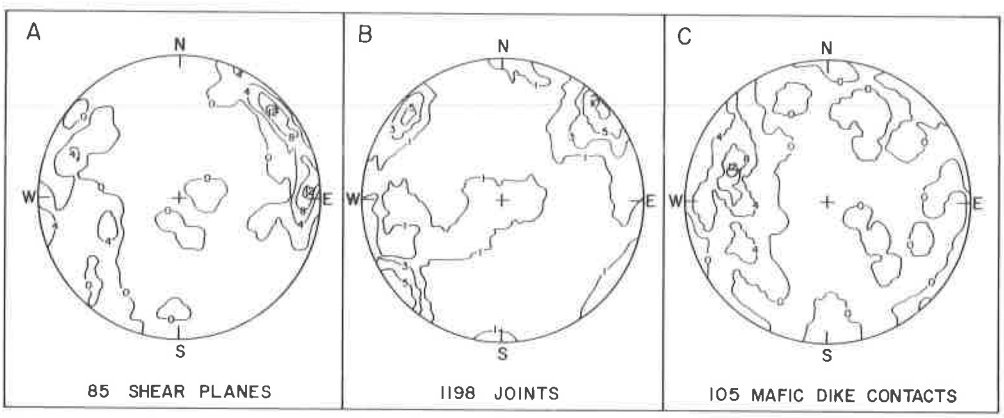

Figure 9. Equal area projections (lower hemisphere) of poles to planes. A. Shear planes. B. Joints. C. Contacts of mafic dikes.

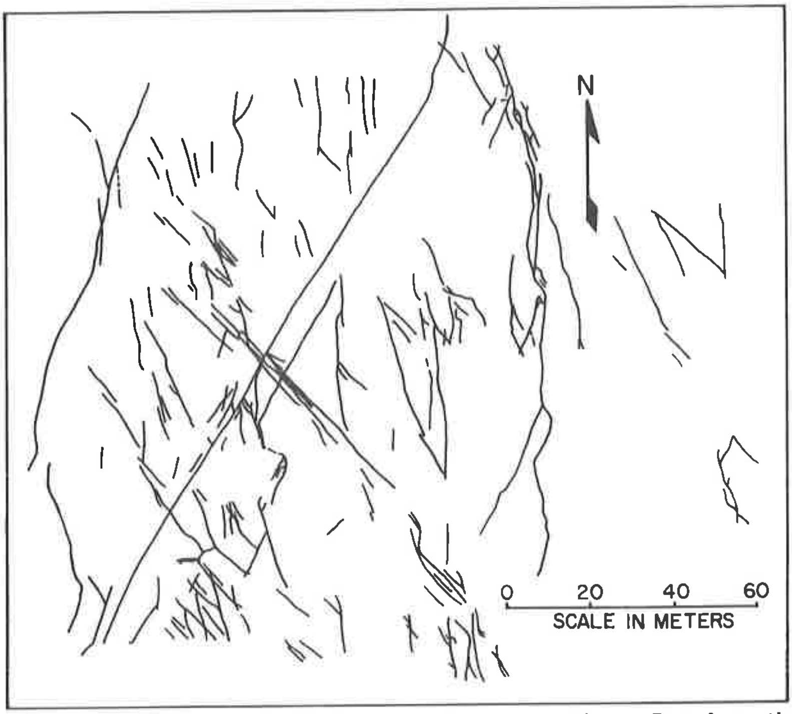

Figure 10. Outcrop pattern of all shears in the excavation. Based on the detailed geologic maps.

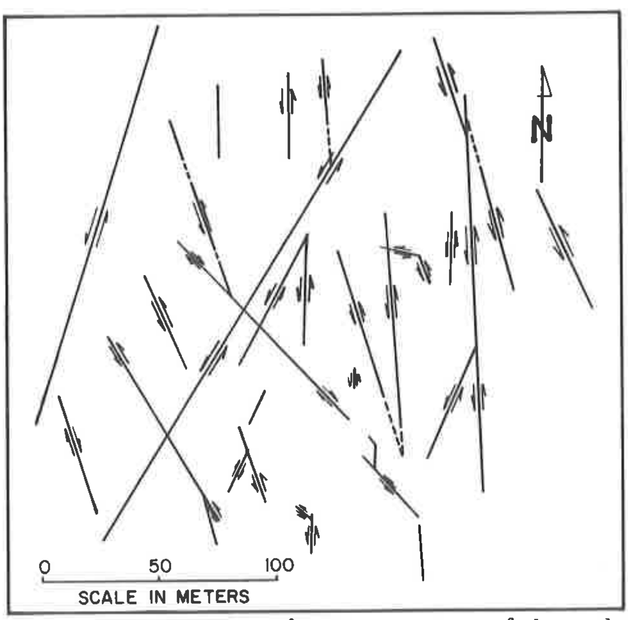

Figure 11. Straight-line generalization of outcrop pattern of shear planes on Figure 10. Direction of apparent horizontal offset indicated, where evident.

planes. The 188 shear planes in the excavation were examined and their orientations were plotted on rose diagrams (Figure 12). When considering all 188 shear planes, the northwest-trending set is dominant, about 38 percent versus about 20 percent for the north-trending set (Figure 12A). When only those shears less than 23 m long are examined, the northwest-trending set is by far the dominant one, 42 percent versus 17 percent for the north-trending set (Figure 12B). The north-trending set is dominant with 29 percent when only those shears longer than 23 m are considered (Figure 12C). The north-northeast set, barely visible in Figures 12A and 12B, is more pronounced in Figure 12C, though still subordinate.

Rose diagrams also were used to examine the length of shear planes in a given orientation (Figure 13). The 188 shear planes have a total length of 3191 m. The north and northwest sets are dominant with 27 and 26 percent of the total length, respectively (Figure 13A). The 38 shear planes that are longer than 23 m have a total length of 1682 m. The north-trending set is dominant among shears above that minimum length criterion with 31 percent (Figure 13B). The north-northeast set is much more pronounced here than it is when no minimum length is considered. Considering only the 14 shear planes longer than 38 m, a total length of 1030 m, the north and the north-northeast sets are dominant with 27 and 23 percent, respectively (Figure 13C).

From a qualitative view (in the field), the largest apparent horizontal offsets are left lateral and occur along the north and north-northeast sets. The dominance of left-lateral offset also is apparent from a quantitative (frequency of occurrence) examination of the data. The 188 shear planes are shown on a rose diagram along with the sense of movement associated with each shear direction (Figure 14). The only one of the four shear directions that displays a dominant right-lateral sense of movement is the northwest-trending set.

Wrench Fault Analogue

There is a geometric similarity between the shear pattern at the site and a theoretical wrench fault pattern as presented by Moody and Hill (1956). The four dominant shear directions at the site may be associated with sets of first-order faults, thus making possible the inference of primary stress directions (Figure 15). It could be assumed that the north direction is the primary, first-order, left-lateral component and the northwest direction is the complementary, first-order, right-lateral component of a wrench fault system (shown with solid lines on Figure 15). Then the azimuth (alpha) of the primary stress direction would be 340 degrees, and the angle (beta) between the stress direction and each of the primary faults would be 20 degrees. It could also be assumed that the north-northeast direction and the north-northwest direction, shown with dashed lines on Figure 15, are similar first-order components of a wrench fault system. For that system, alpha would be 2-1/2 degrees and beta would be 22-1/2 degrees. The theoretical fracture pattern generated by those two stress fields can be correlated with the generalized pattern of shears in the excavation (Figure 16).

When a rock mass responds to stress by failure along a plane, the stress field in the mass is altered. We can speculate on whether the north-oriented stress direction is a response to the north-northwest stress or vice versa. It is reasonable to believe that they are penecontemporaneous, because both the north-northeast and the north-oriented shears parallel major orientations of mafic dikes.

Wrench fault systems such as that described by Moody and Hill (1956) are based on brittle conditions, a constant stress field, material that is homogeneous and isotropic, and contemporaneousness of fracture formation. When faulting was initiated at this site, conditions were not purely brittle as indicated by the branching, anastomosing nature of the north- and north-northeast-trending shears and their close association with mafic dikes. The presence of the various dike materials indicates that conditions during faulting were neither homogeneous nor isotropic. The more linear nature of the northwest-trending shears suggests that they may have developed after the north and north-northeast sets and that fracture formation was not contemporaneous as would be necessary in Moody and Hill's (1956) model. Other evidence, discussed earlier, indicates faulting at this site occurred intermittently over a long

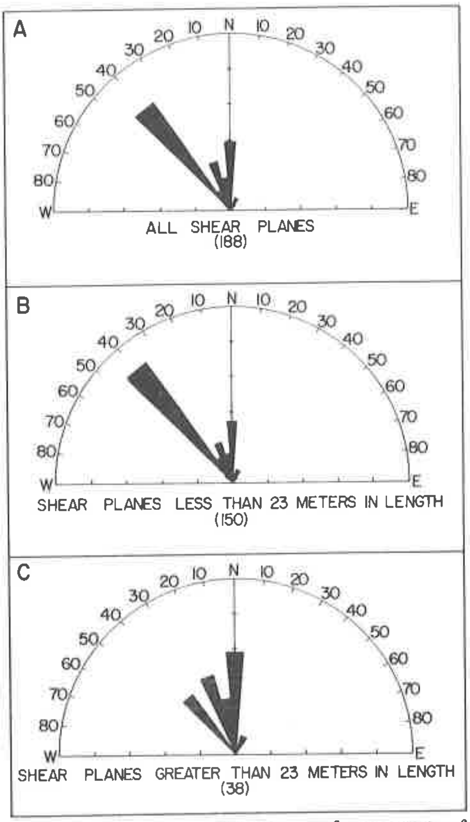

Figure 12. Rose diagrams of percent of total number of shear planes based on examination of field maps of the excavation. Ten percent increments are marked. A. All shear planes. B. Shear planes less than 23 m in length. C. Shear planes greater than 23 m in length.

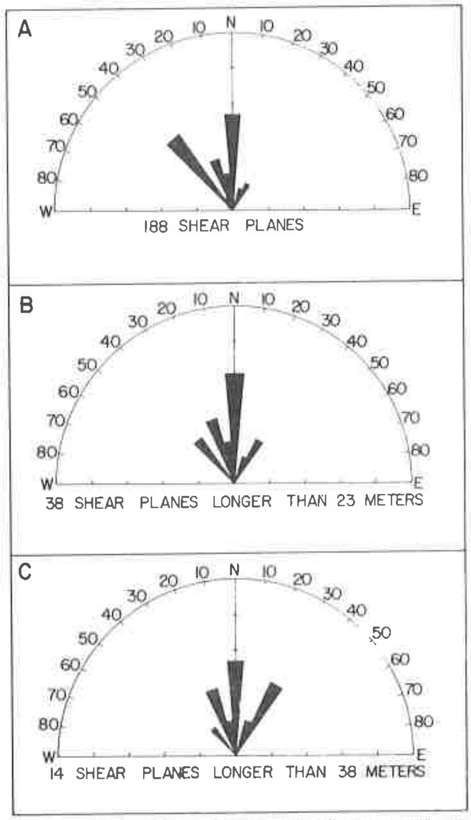

Figure 13. Rose diagrams of percentage of length of shear planes within groups defined by length. Data based on examination of field maps of the excavation. Ten percent increments are marked. A. All shear planes (188) - total length 3191 m. B. Thirty-eight shear planes longer than 23 m - total length 1682 m. C. Fourteen shear planes longer than 38 m - total length 1030 m.

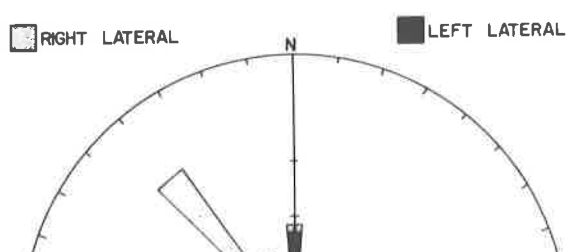

Figure 14. Rose diagram of all shear planes (ref. Fig. 12A) showing distribution of sense of movement (apparent horizontal offset) with shear direction. Ten percent increments are marked.

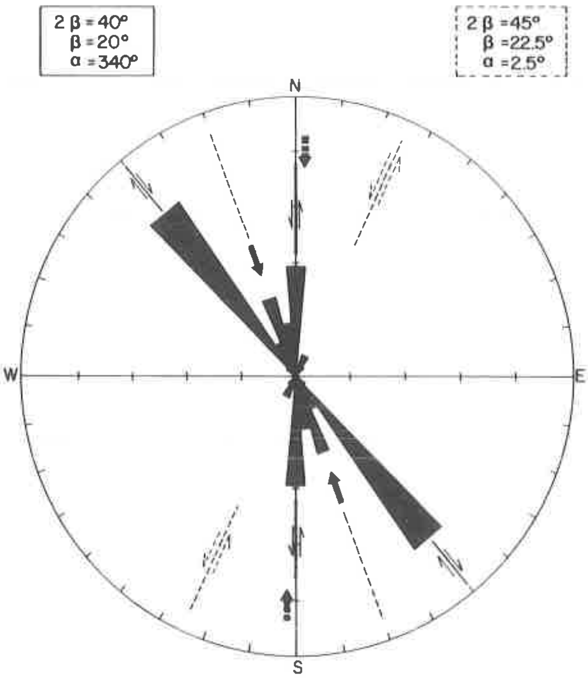

Figure 15. Rose diagram of all shear planes (ref. Fig. 12A) showing associated theoretical stress directions and first order fractures. Ten percent increments are marked. The two possible interpretations as described in the text are shown with solid and dashed lines.

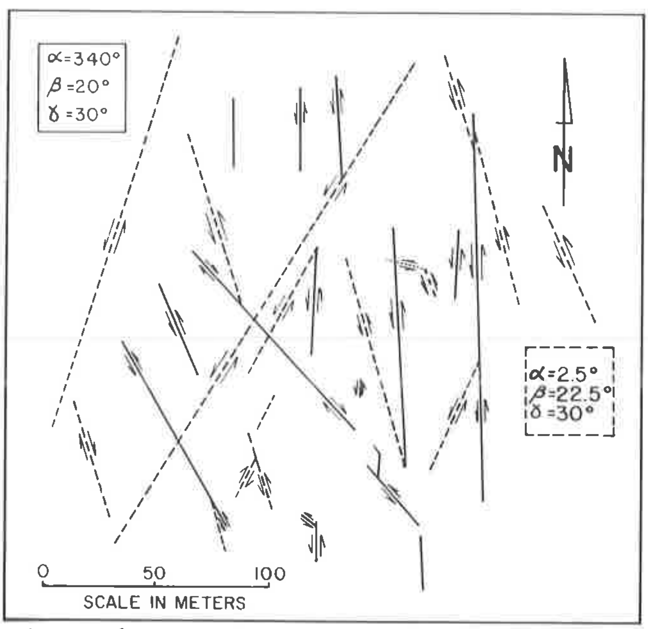

Figure 16. Correlation of straight-line outcrop pattern of shear planes (Figure 11) with first and second order shear directions generated by the two stress systems (Figure 15). Note that only one of the shears (dotted line) cannot be correlated with the theoretical pattern; it may be a third order shear.

time interval, from about 500 to 200 million years ago.

It is concluded that the initial shearing occurred primarily along mafic dikes under ductile conditions (possibly during intrusion and/or amphibolite metamorphism) and that these shear directions (north and north-northeast), as much as the prevailing stress field, controlled the directions of later, semi-brittle to brittle deformation.

DISCUSSION OF GEOLOGIC HISTORY

Geologic features existing at the site were produced by heating and stress events accompanied at times by hydrothermal activity. The apparent sequence in which these events occurred is revealed by analysis of features observed in exposures of rock and saprolite and in thin sections. Figure 17 illustrates our interpretation of the geologic history of the site and possible correlations of local events to regional orogenic events.

Rock at the site is essentially adamellite containing pegmatite, aplite and mafic dikes. Adamellite crystallized 532±15 million years ago. The earliest post-adamellite rocks are pegmatite and aplite dikes. These early dikes were cut by and often offset along the somewhat later mafic dikes. Field observations suggest that the adamellite was still partly plastic when the mafic dikes were injected. Thus, most rocks at the site apparently crystallized penecontemporaneously about 500 million years ago. Plutonism at this site was pre-Taconic and is correlative with late Precambrian-Cambrian volcanic-plutonic activities in the southern Appalachians as described by Butler and Ragland (1969), Fullagar (1971), and Fullagar and Butler (1979).

Evidence of the earliest (pre-metamorphic) faulting at the site consists of pegmatite dikes that are offset where they are cut by mafic dikes. The timing of this faulting is interpreted to be late magmatic. This episode might be related to onset of the Taconic orogeny, but more likely is an older, local deformation related to the plutonic activity. There is no direct evidence that the site was affected by Taconic deformation in the interval about 500 to 450 million years ago.

Regional metamorphism affected the site during the interval approximately 420 to 300 million years ago, comparable to the metamorphic interval observed by Bobyarchick and Glover (1979) in the Piedmont of Virginia. Metamorphism reached its peak at least 380 million years ago with the development of amphibolite-grade mineral assemblages and foliation of variable intensity. Minerals typical of the greenschist facies, essentially restricted to shear and breccia zones, are superimposed on amphibolite-facies mineral assemblages. Quartz-epidote veinlets cut across foliation and indicate the occurrence of hydrothermal activity during this episode. Greenschist metamorphism ended about 300 million years ago.

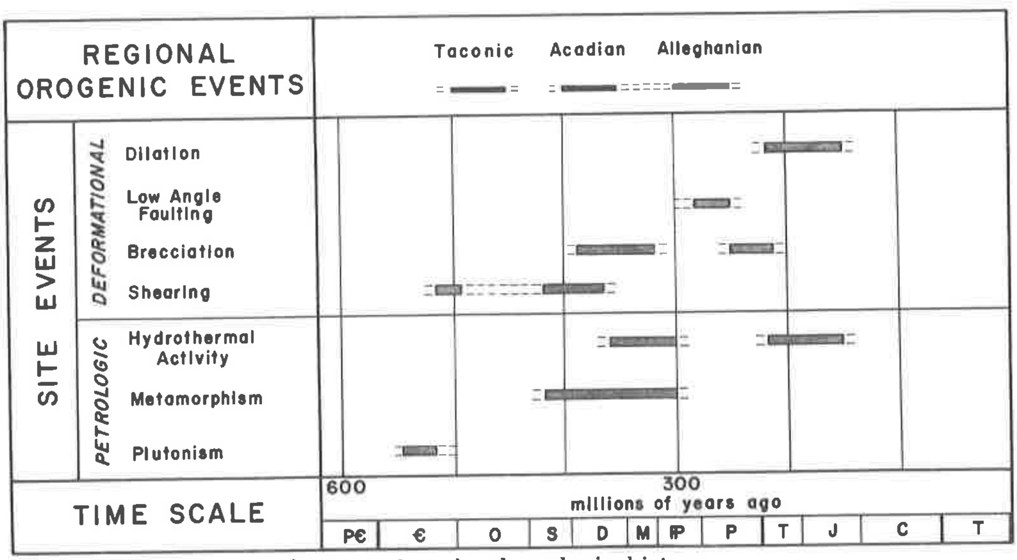

Figure 17. Summary of site and regional geologic history.

Post-crystallization (syn-metamorphic) deformation related to compressional events resulted in two major types of faulting: shearing and brecciation. The orientations of the primary first-order faults during these events were controlled by the orientations of the mafic dikes and an approximately north-south stress field. Secondary faults were controlled by the adjustment of the rock mass to the stress field, perhaps as parts of a wrench fault system. Shearing is closely related to amphibolite-grade metamorphism. Deformation continued as brecciation, in part superimposed on earlier shearing. Shear and breccia zones are healed by greenschist metamorphism, indicating that major deformation ended prior to 300 million years ago. The episodes of metamorphism and deformation described above are consistent with the Acadian orogeny and fall within the temporal range of deformation in the southern Appalachians discussed by Hatcher and Odom (1980).

Following deformation and metamorphism, minor faulting occurred along low-angle planes. These low-angle faults may be related to an episode of thrusting that was part of Alleghanian (Hercynian) deformation (Kish and others, 1978). Low-angle faults are truncated by fractures filled with hydrothermal minerals. Following low-angle faulting, brecciation occurred along paths of the earlier shearing and brecciation. These steeply dipping planes are healed and filled with hydrothermal minerals.

The final deformational and petrologic events to affect the site are dilation fracturing and extensive development of prehnite, calcite and zeolite veins. This dilation fracturing and hydrothermal activity may correlate with widespread diabase dike emplacement and continental rifting that affected eastern North America during Mesozoic time (Butler, 1977). A potassium-argon analysis of laumontite yielded an age of 86 ± 30 million years. The true age of this laumontite is inferred to be about 150 million years or older based on its potential for loss of radiogenic argon and its common association with Mesozoic diabase dikes (although diabase is not present at the site).

With the exception of a general lack of evidence for Taconic deformation, the geologic history of this site is consistent with deformational histories presented in the regional models of other workers.

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LATE-QUATERNARY VALLEY-FILL DEPOSITS IN NORTH-CENTRAL MISSISSIPPI

By

E.H. Grissinger, J.B. Murphey, and W.C. Little
USDA Sedimentation Laboratory
United States Department of Agriculture
P.O. Box 1157
Oxford, MS 38655

ABSTRACT

Six valley-fill deposits have been identified in the north-central Mississippi study area. These deposits occupy consistent relative positions in the landscape. Abundant well preserved wood and other organic detritus occur in association with these deposits. The ^{14}C ages of these organics establish the age of the deposits as late Quaternary. The age frequencies are comparable with sedimentary age relations for diverse areas and are consistent with botanic evidence of paleoclimatic conditions, indicating regional control of the sedimentary sytems. This finding, together with the nature of the valley-fill deposits in the study area, suggests that paleoclimate has been the dominant control of depositional processes.

The nature of these deposits significantly influences present-day channel stability and long-term response of the channel system to changing flow regimes. Individual valley-fill deposits exhibit innate types of channel bed and/or bank failure, thereby imposing controls on the present channel system.

INTRODUCTION

Most streams of north-central Mississippi flow to the west, through the bluff area of loess-capped hills of moderate relief, to the relatively flat flood plain of the Mississippi River (Figure 1). Within and immediately east of the loess area, streams have incised into valley alluvium presently identified as undifferentiated Holocene deposits. Many if not most of these streams are presently unstable, and channels have deepened and widened at a rapid rate. This condition is an economic problem not only within the loessial hills but it also contributes to flood-producing channel aggradation and plugging in the Mississippi River Valley.

As an initial phase of a comprehensive study of stream channel instability, we visually inspected many trenched stream channels east of the bluff line. Bed and bank materials were not uniform for any individual stream channel. We observed, however, that most materials could be grouped into one of several units. Each of these individual units had a consistent appearance throughout the study area and each possessed distinctive properties facilitating differentiation between units. Additionally, each unit occupied a consistent relative position in all watersheds. These observations indicated that the presently undifferentiated flood-plain deposits may include several identifiable units, with each unit having a distinguishable lithology. This paper presents the initial results of our study of these alluvial deposits including (a) the distinguishing properties of each unit, (b) the chronology of the units and (c) the influence of these units on present-day stream bed and bank stability.

VALLEY-FILL DEPOSITS

Six identifiable valley-fill units crop out in channels of the study area. These units contain abundant wood or other organic detritus. Wood identifications and ^{14}C ages are listed by sample locations in Table 1, and sample locations are shown in Figure 1. A frequency histogram of ages less than 13,000 ^{14}C years Before Present (yr BP) is presented in Figure 2a. All ages were calculated using the Libby half-life of 5568 years. No correction has been made for variation in atmospheric ^{14}C concentration. The six units, from oldest to youngest, are (a) consolidated sandstone, (b) bog-type deposit, (c) channel lag deposit, (d) massive silt, (e) meander-belt alluvium and (f) postsettlement alluvium. An idealized section for these units is presented in Figure 3.

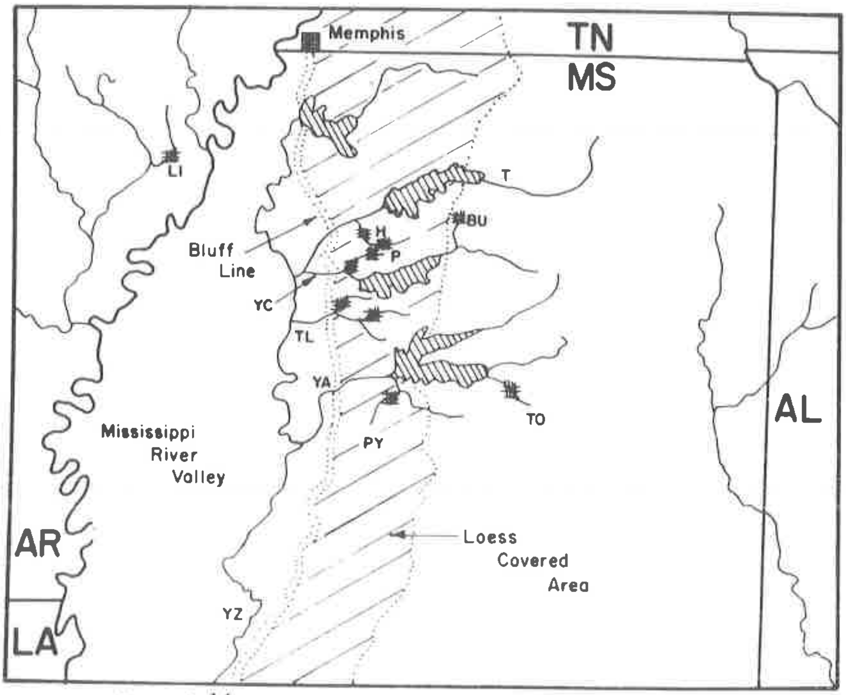

Figure 1. Locations of ^{14}C samples in north-central Mississippi: BU is Burney Branch; H is Hotophia Creek; PY is Perry Creek; P is Peters Creek including Long, Goodwin and Johnson Creek tributaries; T is Tallahatchie River; TL is Tillatoba Creek; TO is Topashaw Creek; YA is Yalobusha River; YC is Yocona River and YZ is Yazoo River. The sample from Lick Creek, Arkansas (identified as LI) is outside of the main area of study. General sample locations are identified by hatch lines.

Consolidated Sandstone

Consolidated sandstones are present in many watersheds. These sandstones are usually cross-bedded and frequently contain gravel. At three locations the sandstone contained wood, all of which was older than 40,000 ^{14}C yr BP (samples I-10,198; I-10,388; and I-10,617). Many of these exposures are enriched with iron and frequently contain iron-replaced wood. Four samples of this iron-replaced wood have been identified as relict of some type of conifer [1/] but the lack of cellular detail prevented further identification. At one location, carbonized wood (sample I-10,617) was recovered from the center of an outcrop of unusually high iron content. Paleomagnetic analysis was attempted at this and nearby exposures, and the direction of magnetization was found to be scattered about the present normal field direction.[1/] Although the paleomagnetic data should be considered preliminary, they do indicate that the time of iron enrichment was materialy younger than the > 40,000 yr BP wood age.

Outcrops of the consolidated sandstone are usually limited in size, rarely exceeding several tens of meters in horizontal distances. The sandstone is typically truncated and outcrops are disconformably overlain by fluvial valley-fill deposits with a maximum age of 12,050±180 yr BP (sample I-10,580, Hotophia Creek). We have not found any outcrops containing wood of boreal species and have not found any outcrops containing datable wood older than 12,050±180 yr BP (Table 1). Additionally, extensive exploratory drilling in the upper Peters Creek watershed (Grissinger and others, 1981) did not encounter any such wood deposits nor any fine-textured deposits which indicated a buried flood plain. A total of 50 exploratory holes were drilled in valleys

1/ Fossil wood identification was performed by S. Manchester, Indiana University. Paleomagnetic analyses were performed by S. Bressler, U.S. Geol. Surv., Flagstaff, AZ.

Table 1. ^{14}C dates, wood identification, deposit type and general location(a).

Type of Deposit(b) Sample	Depth m	Elevation(c) m	Wood Identification	^{14}C Age yr BP±1s
			Burney, Lafayette	
Postsettlement alluvium(d)				
I-10,449	2.7	95	Quercus sp.	<190
Meander-belt alluvium				
I-10,450	3.0	94	Quercus sp.	395±75
I-10,214	3.0	94	Castanea dentata	380±80
			Goodwin, Panola	
Meander-belt alluvium				
I-10,452	1.8	72	Populus sp.	250±75
I-10,215	2.7	71	Liquidambar styraciflua	330±80
Channel sand, below M-BA				
I-10,395	5.7	68	Liriodendron tulipifera	4,545±95
Channel lag, below MS(e)				
I-10,200	5.8	68	Ulmus rubra	9,460±140
Bog-type, below MS				
I-10,576	3.7	71	Juglans sp.	9,750±140
			Hotophia, Panola	
Postsettlement alluvium				
I-9,913	1.8	73	Liquidambar styraciflua	<185
Meander-belt alluvium				
I-10,385	3.7	71	Prunus sp.	1,450±85
I-10,386	4.6	70	NI(f)	1,820±85
Channel sand, below M-BA				
I-9,914	4.6	71	Taxodium distichum	5,600±110
Channel lag, below MS(e)				
I-10,580	8.5	79	NI	12,050±180
Bog-type, below MS				
I-10,579	6.4	82	Ulmus sp.	11,880±170
I-10,615	6.4	82	Acer sp.	11,330±170
I-10,616	6.4	82	Acer sp.	11,660±170
			Johnson, Panola	
Postsettlement alluvium				
I-10,939	3.2	82	Quercus sp.	>190
Meander-belt alluvium				
I-10,577	2.0	71	Castanea dentata	575±75
I-10,578	3.8	85	Sassafras albidum	955±80
I-10,695	4.0	85	Acer sp.	470±80
I-10,940	4.0	81	NI	2,980±85
I-10,953	4.3	85	Cercis canadensis	225±75
I-10,957	4.0	84	Carya sp.	250±75
I-10,964	4.0	84	Quercus sp.	325±75
Bog-type, below MS				
I-10,942	3.7	82	Betula sp.	11,070±160
I-10,943	4.6	83	Quercus sp.	10,110±150
I-10,954	4.3	86	Ulmus rubra	9,485±160
I-10,955	4.1	86	Quercus sp.	8,550±140
I-10,956	4.0	86	Quercus sp.	8,830±145
I-10,965	4.1	83	Fagus grandifolia	8,740±145
Channel lag, below MS				
I-10,696	4.0	87	Platanus occidentalis	9,640±155
I-10,941	3.4	83	Hardwood	10,030±150

Type of Deposit(b) Sample	Depth m	Elevation(c) m	Wood Identification	^{14}C Age yr BP±1s
			Lick, Phillips	
Bog-type, below MS				
I-10,197	5.5	44	Fagus grandifolia	9,510±140
			Long, Panola	
Postsettlement alluvium				
I-10,412	3.7	78	Quercus sp.	<195
Meander-belt alluvium				
I-10,398	5.2	65	Juglans nigra	2,365±85
I-10,211	6.4	64	Liquidambar styraciflua	190±80
I-10,618	3.0	68	Ulmus americana	210±75
I-10,619	5.0	66	Castanea dentata	405±75
I-10,691	5.0	66	Cercis canadensis	455±80
Consolidated sandstone				
I-10,617	6.7	65	Conifer	>40,000
			Perry, Grenada	
Meander-belt alluvium				
I-10,210	4.7	51	Catalpa sp.	1,260±80
I-10,397	2.7	74	Quercus sp.	270±80
Channel sand, below M-BA				
I-10,396	4.6	67	Sassafras albidum	4,830±100
			North Fork Tillatoba, Tallahatchie	
Meander-belt alluvium				
I-9,915	2.0	64	NI	990±80
C-5	2.1	64	Quercus sp.	ND(f)
C-7	1.8	65	Quercus sp.	ND
I-10,451	4.3	61	Liriodendron tulipifera	305±75
C-42	4.3	61	Acer sp.	ND
C-43	4.3	61	Platanus occidentalis	ND
Channel sand, below M-BA				
I-10,694	4.9	59	Fraxinus sp.	6,120±115
			Middle Fork Tillatoba, Tallahatchie	
Meander-belt alluvium				
I-10,415	5.2	62	Quercus sp.	1,315±80
I-10,416	7.0	60	Platanus occidentalis	2,405±90
Channel sand, below M-BA				
I-10,212	5.8	61	Celtis occidentalis	4,050±105
Channel lag, below MS(e)				
I-10,413	8.8	58	Ulmus americana	10,550±150
I-10,414	8.8	58	Betula nigra	10,280±150
I-10,447	7.0	60	Fagus grandifolia	9,840±150
I-10,213	10.7	56	Platanus occidentalis	10,200±150
Bog-type, below MS(e)				
I-10,448	10.8	56	NI	10,760±150
			Topashaw, Calhoun	
Bog-type, below MS				
I-9,917	5.6	80	Quercus sp.	10,080±145
I-10,387	5.6	80	Quercus sp.	10,730±150
Channel lag, below MS				
I-10,199	6.5	80	Quercus sp.	9,410±140
I-10,389	6.5	80	Quercus sp.	9,460±140
Consolidated sandstone				
I-10,198	7.2	76	Platanus occidentalis	>40,000
I-10,388	7.3	76	Hardwood	>40,000

(a) ^{14}C analyses were performed by Teledyne and samples are identified by Teledyne sample numbers. Undated samples are identified by USDA Sedimentation Laboratory sample numbers. All sample locations are in Mississippi except Lick Creek which is in Arkansas. All wood samples were identified by R. B. Miller, U.S. For. Serv., Madison, WI. The fossil-wood sample (I-10, 617) was identified by S. Manchester, Indiana University. (b) PSA is postsettlement alluvium, M-BA is meander-belt alluvium, MS is massive silt. (c) Mean sea level. (d) Possibly meander-belt alluvium. (e) The massive silt was not observed at these sample sites. The relative stratigraphic positions were established from nearby exposures. (f) NI is not identified, ND is not dated.

of this watershed.

Delcourt and Delcourt (1977) reported comparable ages of 12,740±300 to 12,600±275 yr BP for wood samples from the base of a fluviatile terrace in southeastern Louisiana. However, they reported both *Picea glauca* (white spruce) and *Larix laricina* (tamarack) fragments in this deposit. A slightly older maximum age, 14,650±500 yr BP has been reported for a fluvial deposit along the Tombigbee Waterway (Curren and others, 1976) southeast of our study area. More significantly, Delcourt and others (1980) have reported ages ranging from 17,200 to 22,300 yr BP for deposits within 2 to 3 m of the present surface of Nonconnah Creek Valley immediately south of Memphis, Tennessee and about 90 km north-northwest of Peters Creek (Figure 1). The arboreal pollen assemblage of these deposits was also dominated by *Picea* with low concentrations of *Abies* (fir) and *Larix*. They reported that macrofossils of *Picea glauca* were abundant.

Several of the previously discussed findings about the valley-fill sequence in our study area indicate a period of valley erosion prior to 12,050±180 yr BP. These findings include (a) the absence of datable organics older than 12,050±180 yr BP, (b) the absence of boreal species in our samples, (c) the excessive depth of several of the 10,000 yr BP deposits (for examples I-10,448 at 10.8 m and I-10,213 at 10.7 m below ground surface elevation, Middle Fork Tillatoba) and (d) the disconformable contact between the younger valley-fill deposits and the > 40,000 yr BP consolidated sandstone. Such valley erosion is logical, resulting from the interaction of post-glacial pluvial conditions (Fairbridge; 1970, 1972, 1976) with low (relative to present) base-level controls. Although absolute base-level controls are unknown, relative controls can be estimated from sea level and flood-plain elevation changes. Sea levels were about 30 m below present about 12,000 yr BP (Blackwelder and others, 1979) and flood-plain elevations for the Mississippi River were probably about 6 to 8 m below present in northern Mississippi (Saucier, 1974). Flood-plain elevations prior to 12,000 yr BP are poorly defined. This suggested period of valley erosion in north-central Mississippi is comparable to a period of erosion documented by Ruhe (1969) in Iowa.

Bog-Type and Channel Lag Deposits

All samples of the bog-type and channel lag deposits have ages defined by the frequency mode about 10,000 yr BP (Figure 2a). The age span of this mode is interesting; it is generally synchronous with the time man first appeared in the lower Mississippi River Valley (Saucier, 1974) and with the time of excessive Pleistocene generic extinction (Grayson, 1977). We interpret this period as transitional between the preceding period of valley erosion and the subsequent period of deposition of fine-grained materials (see next section).

Although these two units have similar ages, they have different lithologies. As defined herein, bog-type sediments are fine-grained organic-rich materials deposited from low-energy fluvial systems. Most of these deposits appear to have formed either in channel cutoffs or in separation zones downstream from bar deposits. Channel lag materials are coarse-grained, frequently cross-bedded materials deposited from high-energy fluvial systems. Gravel is common where not limited by source availability. Both deposits contain abundant organics. In general, the organic debris in the lag deposits is relatively coarse, ranging up to 100 cm diameter logs. Bog-type organics include leaves, twigs, various nuts, and scattered stumps and logs similar in size to those found in the lag deposit. There is little evidence of abrasion or aerobic decomposition in either deposit, indicating rapid burial. Heartwood, alburnum and bark are usually well preserved and the cellular structure of woody tissue is intact. In addition, acorns with caps attached; complete leaves; and walnuts, butternuts, and hickory nuts, frequently complete with husks have been seen in the bog-type deposits.

The 3,500 ^{14}C yr range of individual ages within the 10,000 yr BP mode (sample I-10,580, Hotophia Creek at 12,050±180 to sample I-10,955, Johnson Creek at 8,550±140 yr BP, Table 1) reflects both within-watershed variation and between-watershed variation. The eight Johnson Creek samples within this mode had an age range of 2,500 ^{14}C yr, suggesting a time of relatively constant base level. Depth below present ground surface for these eight samples ranged from 3.4. to 4.6 m. Within this mode, the four Hotophia Creek samples were the oldest and were an average of 660 ^{14}C yr

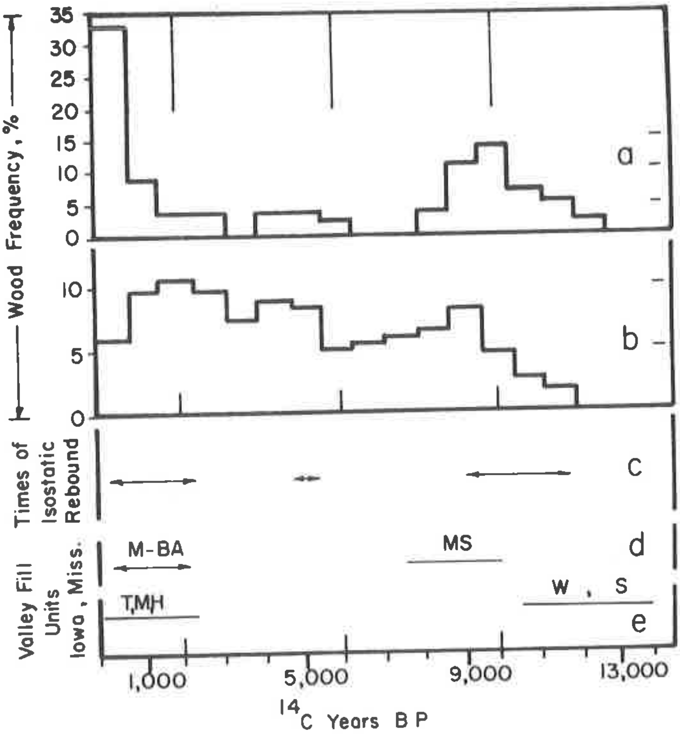

Figure 2. Chronology of valley-fill deposits. (a) Frequency histogram of ^{14}C dates for 57 samples from north-central Mississippi (source: this paper). (b) Frequency histogram for 815 ^{14}C dates selected from the journal Radiocarbon [source: Wendland and Bryson (1974)]. (c) Times of paleoclimate-controlled isostatic rebound in the Great Lakes area [source: Flint (1957)]. (d) Deposit ages in north-central Mississippi; M-BA = meander-belt alluvium, MS = massive silt (source: this paper). (e) Valley-fill ages in Iowa; T = Turton, M = Mullenix, H = Hatcher, W = Watkins, S = Soetmelk [source: Daniels and Jordan (1966) and Ruhe (1969)].

older than the oldest sample from any other site. This greater age for the Hotophia Creek samples is possibly related to the relative position of this creek in the drainage net of the Yazoo River system (Figure 1). Hotophia Creek is a tributary of the Tallahatchie River, the major tributary within the Yazoo River system, and joins the Tallahatchie near the present Mississippi River Valley. As the major tributary within the drainage net of the Yazoo River system, the Tallahatchie River and its tributaries probably would respond more rapidly to Mississippi River Valley base-level controls. The average ages for samples in this early-Holocene mode for the four main watersheds ranged from 9,600 to 11,700 yr BP (Table 2). Qualitatively, these average ages for tributaries of bluff line watersheds increase with increasing watershed size (Tallahatchie > Yalobusha > Yocona). Tillatoba, however, is a primary bluff line watershed (it drains directly into the Mississippi River Valley) and samples from this location have average ages intermediate between tributaries of the Tallahatchie and Yalobusha watersheds. These relations must be considered as preliminary trends because the number of samples is limited. But they do suggest the overall significance of base level as a control of the valley-fill sequence.

Massive Silt and Meander-Belt Alluvium

Both units are relatively fine-grained valley-fill deposits and both are buried beneath postsettlement (historic) alluvium. Each unit has a consistent set of depositional features and a distinctive, characteristic weathering profile. Properties of these two units pertinent to present-day channel bed and bank stability have been

Table 2. Average age of samples in the 10,000 yr BP mode for streams in the Yazoo River system.

Main Watershed(a) / Intermediate Tributary / Tributary	Number of Samples	Average Age ^{14}C yr BP	Distance From Bluff Line(b) km
Tallahatchie			
Hotophia	4	11,700	17
Yalobusha			
Topashaw	4	9,900	77
Yocona			
Peters			
Goodwin	2	9,600	16
Johnson	8	9,600	16
Tillatoba	5	10,300	10

(a) The term main watershed is used to identify those streams which flow from the loess area into the alluvial valley of the Mississippi River. The relative watershed size for these streams is Tallahatchie>Yalobusha>> Yocona>>Tillatoba.

(b) Approximate distance.

produced by both depositional and weathering features. We refer to weathering in the massive silt as paleosol II weathering and that in the meander-belt alluvium as paleosol I. These designations minimize possible confusion involving current soil classification units and associated weathering features. Current soil classification units are materially influenced by the overlying postsettlement alluvium.

Massive Silt: The massive silt deposit does not contain any organics suitable for dating. It immediately overlies bog-type sediments and channel lag deposits which have an age of about 10,000 yr BP, establishing the maximum age of the massive silt. Relict entrenchment into or through the massive silt deposit is common. Five wood samples (I-10,396, Perry Creek; I-10,694, North Fork Tillatoba; I-10,212, Middle Fork Tillatoba; I-9,914, Hotophia Creek; and I-10,395, Goodwin Creek) have been obtained from fill deposits of this relict entrenchment. These woods range in age from 4,050 to 6,120 yr BP and comprise the frequency mode at about 5,000 yr BP (Figure 2a). Fill deposits of this age are common but are not areally extensive. We have not differentiated these materials as a valley-fill unit but refer to them as channel sands (Table 1). They are significant in that they establish the minimum age of the massive silt deposit at about 5,000 yr BP.

The massive silt is a widespread, predominately fine-grained, valley-fill deposit. It is distribued throughout the study area and frequently exceeds 4 m in thickness. The deposit fines upward from a silty sand or sandy silt basal material to a silt, with no observable textural breaks except for occasional small relict channels which appear to be extensions of hillslope tributaries. No large relict channels or oxbows have been observed which would indicate main-channel flow at this time of valley aggradation. Additionally, no pre-existing flood-plain deposits have been identified at an elevation comparable with that of the massive silt. Bedding is rare and has been observed only in the sandier basal material. No bedding characteristic of point bars, natural levee or splay deposits has been observed in this deposit.

Based on the preceding observations, we interpret the massive silt as a low-energy fluvial deposit, possibly associated with periodic inundation resulting from trunk valley aggradation. Similar plugging has been described by Pflug (1969) for tributaries in eastern Brazil, but at a slightly earlier time than that for this deposit. Aeolian materials may have been an additional source for this massive silt deposit. The contact between the massive silt and channel lag deposits is gradational, indicating that the silt is only slightly younger than the underlying deposits (Figure 2d). We interpret this massive silt deposit as the end member in the sequence valley erosion → bog-type or channel lag deposits → massive silt, this sequence representing a continuing decrease in energy resulting from decreasing pluvial activity and rising base-level controls.

Indirect support for this interpretation is contained in the late-Quaternary discharge record of the Mississippi River and in the paleotemperature record. Glacial meltwater flow down the Mississippi River commenced about 17,000 yr BP. This flow increased steadily and peaked about 13,500 yr BP at which time flow diversion occurred to the east. Meltwater flow volume decreased rapidly until about 11,500 yr BP when normal isotopic composition was restored (Kennett and Shackleton, 1975). The paleotemperature record complements the flow record. Temperature recovery started about 17,000 yr BP and was relatively rapid (Harmon and others, 1979). By about 10,000 yr BP, meteoric water was comparable with that of today, based on D/H ratios (Yapp and Epstein, 1977), indicating effectively complete transition from glacial to interglacial conditions by this time. They describe this transition as asynchronous for North America.

An alternate interpretation of this valley-fill sequence is that the silts represent vertical accretion deposits and the channel lag deposits represent lateral accretion deposits, both resultant from meandering stream flow. The vertical and lateral accretion deposits would generally be equivalent in age using this alternate interpretation. As described previously, however, this deposit contains none of the bedding features typical of deposits associated with meandering stream flow and no relict channels or oxbows. We therefore believe this alternate interpretation is unlikely.

The paleosol II weathering profile formed on the massive silt is distinctive. This paleosol has a thick A_2 and a dense B_2 horizon, both unique to paleosol II. Gray is the dominant color in the upper part of the profile. Iron and manganese stains and concretions are present in the lower B. The B_2 horizon has a well developed polygonal structure, with seams often wider than 2 cm (Figure 4). This unit is relatively infertile and restricts the vertical movement of water. Vegetative cover is rare on outcrops.

Meander-Belt Alluvium: A major entrenchment of streams into the massive silt began about 3000 yr BP (sample I-10,940, Johnson Creek). This time of entrenchment agrees with the paleoclimatic interpretration of Wendland and Bryson (1974) who reported a major botanic discontinuity at 2760 yr BP associated with increased rainfall. The distribution of ages in the youngest frequency mode (Figure 2a) indicates that fluvial activity in our study area was relatively minor until about 1,500 yr BP, whereupon it increased gradually and continued until about 1800 A.D. (the minimum age detectable by ^{14}C analysis). Peak activity occurred within the last 800 yr BP. (Again, the sample size and collection procedure does not preclude sample bias and this age distribution within the youngest mode should be considered a trend.) This finding is in general agreement with observations by Hilgard (1860) who noted that most streams in this part of Mississippi had ". . .cut more or less into the clayey strata of the Lignite formation. . . ." of Tertiary age (page 306, paragraph 656).

The entrenched streams apparently meandered across the flood plains, eroding older materials and depositing the unit identified as meander-belt alluvium in typical meander patterns. These materials are typically vertical accretion overlying lateral accretion deposits with occasional oxbow deposits of layered fines.2/ These two types of deposits have not been separated. Wood is scattered throughout this deposit but is usually less well preseved than older wood. This state of wood preservation is probably due to the greater permeability of this material relative to that of the massive silt. In all cases, bedding is readily observable.

Weathering of the meander-belt alluvium is less intense than that of the massive silt. The paleosol I has an A_1 which varies in thickness from more than 25 cm to only a few cm. The profile is frequently truncated, however, with no observable A. Paleosol I has no A_2 horizon, no B_2 horizon and no polygonal structure. This profile development indicates a period of landscape stability and suggests prairie-type vegetative cover. Iron diffusion halos are usually present in the subsoil but are

2/ Measurements of secular variation have been attempted for varve-like oxbow deposits. Preliminary results show declination 10^{o} to 30^{o} east of North, suggesting secular variation may be a useful correlation tool in the absence of datable wood samples. These determinations were made by S. Bressler, U.S. Geol. Surv., Flagstaff, AZ.

typically small. This paleosol is relatively fertile and well drained.

These findings are generally comparable with observations by Happ and others (1940). Both studies identified a buried, prairie-type soil profile that was truncated in places. Happ and others further stated that prairies were reported in the upper parts of tributary valleys in the General Land Office surveys of 1834-36. In addition, Happ and others (1940, pages 23-24) suggested as one possibility that ". . .lateral migration was more important than at present, so that the channel shifted from side to side. . . ." depositing vertical accretion over lateral accretion materials that together formed this youngest presettlement valley-fill deposit. They also stated that during this time of meandering, the valleys had been ". . .partially reexcavated and graded down to a lower level. . . ." but that evidence was not sufficient to establish if valley excavation was in progress at the time of settlement (page 38). This slight entrenchment is illustrated in Figure 3 for our study area and is thought to be reflected in Hilgard's (1860) frequent references to second bottom lands. The ^{14}C dates presented herein indicate that meander-belt development was active immediately prior to settlement. The fluvial system was evidently a stable meandering system and such a system continuously erodes older deposits and builds new deposits while maintaining an otherwise stable channel. The frequent reports of stream navigability during early settlment times are thus compatible with this scenario.

Postsettlement Alluvium

Postsettlement alluvium (PSA), produced in historic times largely by human activities, caps almost all flood-plain surfaces. This material is frequently less than 1 m thick but may locally exceed a thickness of 3 m. The PSA has well preserved fluvial bedding features. It is unweathered with an Ap horizon directly overlying a C horizon. Iron diffusion halos have not been observed. Although this unit is too young

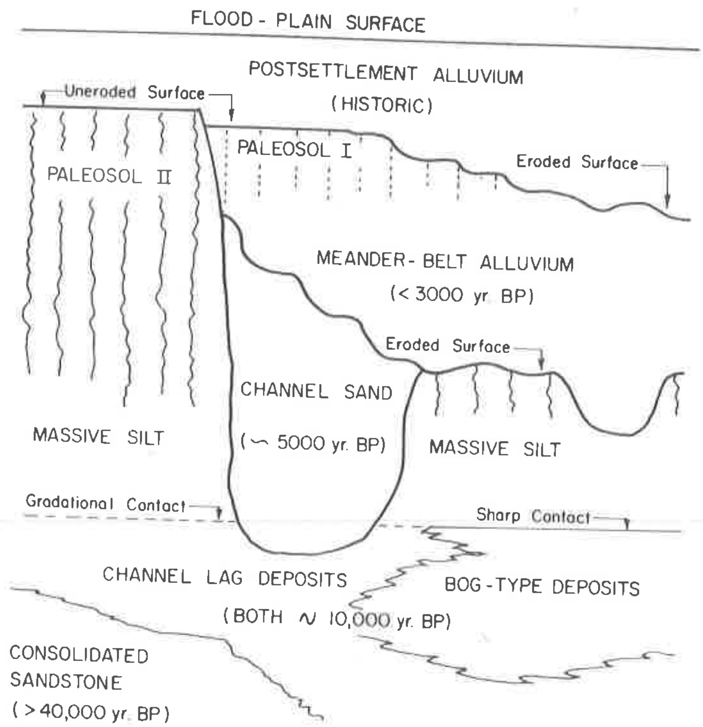

Figure 3. Idealized section of valley-fill units. The stepped surfaces of paleosols I and II illustrate the first and second bottoms described by Hilgard (1860). At most locations these surfaces are now buried under postsettlement alluvium produced during historic time.

to be identified by radiocarbon procedures, it is identifiable in the field by the presence of man-made artifacts above a disconformity. It has been the subject of many reports, including those by Happ and others (1940), Happ (1968, 1970) and Trimble (1974).

DISCUSSION

Paleoclimatic Control

Wendland and Bryson (1974) used 815 dates published in the journal ***Radiocarbon*** ". . .to identify times of large-scale hemispheric discontinuity . . . " based on geologic-botanic discontinuities. Their data base included ^{14}C dates which defined the age of discontinuities within peat beds, pollen profiles, glacial records and sea level stands. A parallel data base of 3,700 dates associated with 155 human cultures was also studied to identify cultural discontinuities. The primary source of this data base was also the journal ***Radiocarbon***. Globally synchronous discontinuities occurred in both data bases and they argued that such results would only be produced if climate was the primary forcing function. Of the seven major geologic-botanic discontinuities which they identified by fitting partial collectives to a multimodal distribution, three are included in the age range of 850 to 2760, one at 5060 and three in the range from 8490 to 10,030 yr BP. Figure 2b is a frequency histogram of their total geologic-botanic data, organized in 800-yr classes to emphasize large-scale trends. The relation between the three groups of discontinuities and the age frequency for their data (Figure 2b) is obvious, as is the similarity of this age frequency with (a) the times of isostatic rebound in the Great Lakes area (Figure 2c) which Flint (1957) attributed to climatic control of glacial ice unloading and (b) the frequency for our data (Figure 2a). The distributions are trimodal with (generally) comparable modal ages. This apparent fit supports Wendland and Bryson's argument that climatic change, and not any specific climate, is the primary forcing function. In relation to our valley-fill data, large-scale climatic changes would logically produce corresponding changes in flow characteristics, and these latter changes would, in turn, produce corresponding changes in the valley-fill deposits. A corollary of this reasoning is that individual valley-fill deposits are chronologic units if the climate is the primary forcing function.

Compared to the data of Wendland and Bryson (Figure 2b) our data (Figure 2a) have a younger average age of the most recent mode, a relatively small frequency of the mode at about 5000 yr BP and a total absence of intermodal ages. We expected this latter difference between the two distributions because of the restricted sample type and source area of our samples. All of our samples were fluvial deposits from tributaries to the lower Mississippi River Valley.

The relative difference in magnitude between the modes near 5000 yr BP appears to fit a paleoclimatic trend of more uniform and drier conditions for our study area relative to the conditions for the areas included in Wendland and Bryson's data base. About 75% of the dates in their data base were European in origin, so their data base was generally equal in origin to that of Starkel (1966). Knox (1975) compared the studies of Starkel (1966) and Wendland and Bryson (1974) and reported that the results were also generally equal. According to Starkel, the Holocene Atlantic period (8150-4950 yr BP) in Europe was a time of climatic optimum. Mean annual temperature was about 2°C warmer than present and the climate was humid. Similar conclusions have been presented by Imbrie and Imbrie (1979) and by Fairbridge (1976) who noted that the northerly shift of the optimum was strongly diachronous. The sub-Boreal period (4950 to 2450 BP) was markedly different (Starkel, 1966). It was warm but rather dry with pronounced fluctuations in humidity.

Climatic conditions in North America were evidently somewhat different. Davis and others (1980) found that the climate of New England from 9000 to 5000 BP was one of maximum warmth or warmth and dryness. During this time, the prairie-forest boundary of the upper Midwest (United States) was east of its present location, indicating warmer and drier conditions than at present (McAndrews, 1967). For both areas, the climatic change to present conditions was gradual. In the Lake Michigan Basin, the post-glacial warming continued until 3500 to 4000 yr BP (Zumberge and Potzer, 1956). This was the warmest and driest period during the Holocene for this area. In Iowa, the change from deciduous trees to nonarboreal species began about

7,000 yr BP with the latter species becoming dominant about 6,000 yr BP (Ruhe, 1969).

Comparable climatic changes have been reported for many areas of the United States. Early-Holocene to mid-Holocene climates either warmer and/or drier than present have been interpreted from vegetation changes established by pollen and in some instances plant macrofossil analyses. These warmer and/or drier conditions persisted in southeastern Missouri from 8700 to 5000 yr BP (King and Allen, 1977); in middle Tennessee, from 8000 to 5000 yr BP (Delcourt, 1979); in northern Florida, from 11,000 to 7,200 yr BP (Watts and Stuiver, 1980); and in southern Florida, from 8000 to 4500 yr BP (Clausen and others, 1979). Climatic recovery to present conditions, if discussed in the preceding references, is usually described as gradual. In the southwestern United States, however, the climate has continued relatively unchanged since early- to mid-Holocene. Bryant (1975) interpreted the pollen record from deposits in Boriack and Hershop Bogs in central Texas to indicate the present climate became established sometime during the Altithermal (from about 7,000 to 4,500 yr BP). Pre-Altithermal climatic conditions were cooler and/or wetter. In the southwestern United States, Van Devender and Spaulding (1979) studied plant macrofossils from packrat middens. They reported that mesophytic species disappeared about 11,000 yr BP and that xeric woodlands disappeared about 8,000 yr BP. Present climatic conditions were established shortly thereafter.

We suggest that this subdued climatic change, in contrast to the relatively rapid transition or times of discontinuity characteristic of the European climatic sequences is responsible for the relatively low magnitude of the mode at about 5,000 yr BP for our data. The rationality for this suggestion is that the magnitude of complex response (Schumm, 1977) will probably vary directly with both the magnitude and rate of change of the forcing function, in this case climatic change.

The younger average age of the most recent mode for our data in relation to the comparable mode for Wendland and Bryson's data is not as easily explained. The peak wood frequency for our data within the last 800 yrs BP is undoubtedly related to cut and fill activity associated with stream meandering. This peak frequency is thus not a result of ". . .times of large-scale hemispheric discontinuity. . . ." as studied by Wendland and Bryson, but rather represents a period of system stability. This stability is further indicated by the paleosol I development. Direct comparison between these two data sets is thus confounded by these system differences.

Although additional difficulties may have been introduced by sampling bias, the distribution of dates within this mode for our data does suggest that this mode may possibly reflect paleoclimatic conditions in the mid-continental United States. Wendland and Bryson noted that the major botanic-geologic discontinuity at 850 yr BP coincided with a distinct change in the Mill Creek culture of Iowa. They hypothesized, on the basis of pollen data, that this change resulted from stronger westerlies. Such a change in circulation patterns would have probably affected large areas of the mid-continental United States, possibly including our study area. The paucity of older ages within this mode is consistent with gradual climatic changes, as discussed previously.

Relations With Other Valley-Fill Deposits

Knox (1975) analyzed 802 dates from the journal ***Radiocarbon***. This data base included dates of flood-plain horizons, terraces, alluvial fans and a few archaeological dates, in aggregate representative of Northern Hemisphere middle latitudes. His results were in good agreement with those of Wendland and Bryson (1974) and Starkel (1966). He concluded that this agreement (a) "supports the concept of global synchroneity of climatic shifts" and (b) "implies that alluvial episodes and the stability of stream channels are strongly influenced by climatic change." Unfortunately, he does not state the percentage of European samples in his data base. This agreement of our results with those of Wendland and Bryson indirectly indicates similar agreement between Knox's analysis and our study. Direct comparison was not attempted.

Knox (1975) also discussed the results of Haynes' (1968) study of alluvial chronologies for the southwestern United States and reported that the discontinuities of the Southwest weakly reflected those presented by Knox, Starkel and Wendland and Bryson. Direct comparison of Haynes' results (1968) with ours is presented in Table 3. Although this comparison is subjective, it does suggest rather synchronous fluvial

Table 3. Comparison with results of Haynes (1968) for the southwest[(a)].

After Haynes		Unit	This Paper	
Unit	^{14}C Age(b)	Description	^{14}C Age(b)	Unit
A	>11,500	fluvial gravels to fines	8,500 to 12,000	bog-type and channel lag
B_2	about 7,000 to 11,000	massive to weakly-bedded silt	slightly younger than about 10,000	massive silt
C_2	about 4,000 to 6,000	channel fill incised into B_2 or massive silt	4,000 to 6,100	channel sand
D	about 2,000 to 4,000	variable	No comparable unit	
E	<1,500	variable, incised into older units	primarily <1,000	meander-belt alluvium

(a) The B_1 and C_1 units of Haynes have been excluded from this comparison. The C_1 is an aeolian unit and the B_1 is described as usually local in extent.

(b) ^{14}C age in yr BP.

activity in these two areas. In Iowa, deposits have been identified with ages equivalent to the meander-belt alluvium. These deposits include the Turton, Mullenix, and Hatcher members of the DeForest formation (Figure 2e) (Daniels and Jordan, 1966; Ruhe, 1969). Although somewhat older, the Watkins and Soetmelk members (Figure 2e) of the same formation appear to be equivalent with the massive silt. Gully cutting and filling in Iowa began about 6,500 yr BP (Ruhe, 1969), between the two times of deposition of DeForest formation members, and this activity may be equivalent with channel incision in north-central Mississippi which started about 6,000 yr BP. Both the Soetmelk member and the massive silt are underlain in places by gravel. Brakenridge (1980) identified a comparable valley-fill sequence in Missouri including (a) aggradation ending about 8,000 yr BP, (b) a period of stability ending about 5,000 yr BP followed by rapid incision and (c) cyclic erosion-aggradation since that time. In an early study, Knox and Johnson (1974) reported no evidence of fluvial adjustment in southwestern Wisconsin to climatic change since about 4,400 yr BP; however, Knox and others (1981) subsequently found two episodes within the time spanning meander-belt deposition in our study area. These times of active erosion and deposition were 3100-1800 yr BP and 1200-800 yr BP. The ages of older valley-fill deposits were synchronous for these two areas.

Ages of buried organics from Georgia (Staheli and others, 1977), Tennessee (Kellberg and Simmons, 1977), Alabama (Curren and others, 1976) and Oklahoma (Gross and others, 1972) generally fall within the modal ages of the frequency distribution (Figure 2a) for our study area. An older sample, dated at 32,000 to 35,000 yr BP, was reported by Kellberg and Simmons (1977) for Tennessee but this material was from a terrace 17 meters above the present flood plain. The numbers of dates and the descriptions of the valley-fill units in these studies are insufficient for a more detailed comparison, however.

Delcourt and others (1980) reported a different valley-fill sequence for Nonconnah Creek in extreme southwestern Tennessee. As previously discussed, they reported organics in fluvial deposits within several meters of the flood-plain surface with ages of 17,200 to 22,300 yr BP. They identified an overlying gray-brown silt layer as loess. No materials equivalent to the massive silt of our study area were described at the Nonconnah site. We suspect these differences resulted from varying base-level controls for these two areas. Nonconnah Creek drains into the Mississippi River through an oxbow lake near Memphis, Tennessee. This location has been a fulcrum of changes in the flood-plain elevation of the Mississippi River over the past 12,000 years, the valley below Memphis aggrading and that north of Memphis degrading. Saucier (1974) describes the flood plain of 12,000 years ago as (a) probably 23 to 24.5 m lower than present south of Baton Rouge, Louisiana; (b) probably 6 to 8 m lower between Vicksburg, Mississippi and Memphis, Tennessee and (c) higher than today north of Memphis. With this rather constant base-level control at Memphis, Nonconnah Creek Valley would not have been subjected to the same degree of stress as was imposed on areas to the south, and such stress must have postdated the youngest valley-fill deposit in the Nonconnah Creek area. This maximum age of 17,200 yr BP for valley erosion is in agreement with the paleotemperature inferences and with general late-Quaternary glacial conditions. Inherently, subsequent valley aggradation would also be minimal.

Channel Stability

The late-Quaternary valley-fill units, together with present hydrologic conditions, control present-day channel stability in this study area. These units exhibit typical types of failure, depending upon their position in the channel bank and/or bed. Bank stability is influenced by postsettlement alluvium, massive silt, meander-belt alluvium and the bog-type and channel lag deposits. Bed (thalweg) stability is influenced by the massive silt, bog-type and channel lag deposits and the consolidated sandstone. Although discussed separately, bank stability cannot be evaluated independently of bed stability; both must be considered for realistic solutions to the massive channel instability problems of the areas bordering the Lower Mississippi River Valley.

Bank Stability: Postsettlement and meander-belt alluvium most frequently occur in an upper-bank position. These materials are well drained, relatively fertile and are usually well vegetated. Scour by high velocity flow is a minor consideration for stability because of the presence of vegetative cover and the infrequency of exposure to high velocity flow. Scour is proportionately more significant for these materials when they are in a lower-bank position. The most frequent erosion problems result from gravity failure accentuated by tension crack development. These tension cracks are vertical and parallel to the channel bank. Their development is undoubtedly related to the relatively unweathered and hence isotropic nature of these deposits.

The polygonal structure (Figure 4) characteristic of paleosol II controls the stability of the massive silt. Although individual blocks (peds) are resistant to channelized flow, the seam materials are only marginally stable. Erosion or weakening of the seam material reduces interped strength, resulting in gravity-induced block failure. We believe this polygonal structure is probably the result of desiccation due to the early- to mid-Holocene temperature maximum. (As previously discussed, evidence for this temperature maximum is widespread but we have no direct evidence for our study area.) The polygonal development was probably accentuated by periodic inundation. Although failure of both fine-textured units is gravity induced, the rate of removal of the slough material is undoubtedly controlled by flow variables. An additional influence of the massive silt on stability results from its low relative permeability. This unit is less permeable than the overlying materials and seep commonly occurs at the interface, further stressing stability of the overlying materials.

The bog-type and channel lag deposits underlie the massive silt. Both are unconsolidated materials of low cohesion and are easily eroded by channelized flow. Channel incision into either of these units invariably results in excessive channel widening due to excessively weak toe conditions.

Bed Stability: Thalweg incision through the massive silt has occurred by headward migration of knickpoints. Two types of migration have been observed, the usual overfall-type failure and a more complex type of failure which is initiated by the development of chutes through the polygonal-structured massive silt. Weak seam materials between individual blocks are winnowed by base flow, isolating individual blocks which are easily displaced by high velocity flow. In Johnson Creek, the rate of knickpoint movement averaged 160 m/yr from 1940 through 1975 (Ethridge, 1979). Thalweg elevations upstream of such knickpoints were generally stable. Exposure of the unconsolidated bog-type and channel lag deposits downstream of the knickpoint, however, resulted in channel widening and changed the flow regimen. Upstream of the knickpoint, channel beds are cohesive. Downstream the channels have sand or gravel

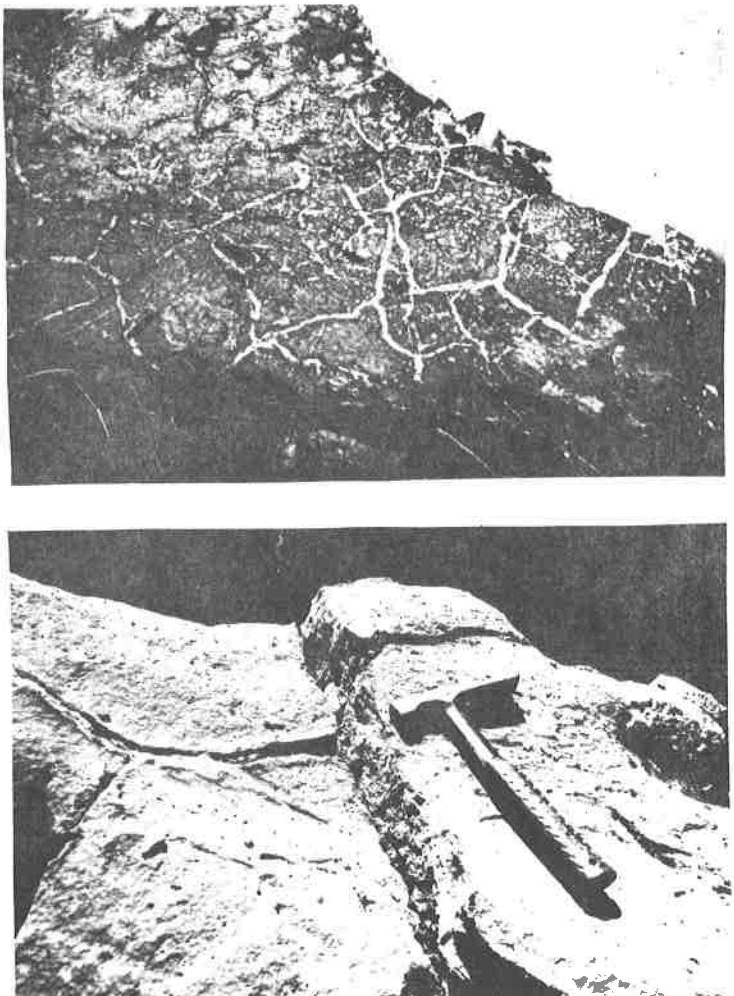

Figure 4. Paleosol II polygonal structure. (a) Unweathered surface. (b) Weathered, disarticulated blocks.

beds and transport processes are dominant. Inherently, the bed stability of these sand-bed channel reaches is primarily dependent upon the available sediment supply and upon the sediment transport properties of the hydraulic system. Consolidated sandstones like those that outcrop in the Goodwin Creek channel limit thalweg lowering and function as local grade controls.

CONCLUSIONS

We have identified six late-Quaternary valley-fill units in our study area. Diagnostic properties are consistent within units and are sufficiently different between units to be usable in the field. The ages of these units have been established by dating 60 organic samples using standard ^{14}C dating procedures. These ages are generally coherent within units but differ between units representing different periods of fluvial activity. The properties of the units, their distributions in the valleys and their ages are consistent with the paleoclimatic record and the record of changes in base-level control (for our study, changes in the elevation of the Mississippi River flood plain).

These units influence channel adjustment to changing flow conditions. The consistency between our results and the base-level and paleoclimatic records indicates that the valley-fill controls for a few channels may be usable as "model" controls for a specific area. We think these findings have potential practical application in the management of our streams and rivers whereby the properties of individual valley-fill units and their distributions could be used in the formulation of optimum channel stabilization design criteria.

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MINERALOGY OF THE TWIGGS CLAY AT ITS TYPE LOCALITY, GEORGIA

By

Orlando A. Rodriguez
Departmento de Suelos
Universidad Centro-Occidental
Aptdo. 400
Barquisimeto, Venezuela

and

J. Anthony Hemphill and Vernon J. Hurst
Geology Department
University of Georgia
Athens, Georgia 30602

ABSTRACT

The Twiggs Clay, a facies of the Upper Eocene Jackson Group in Georgia is composed mainly of smectite and cristobalite-tridymite. The smectite varies in composition and cation site occupancy. It can be categorized as about half montmorillonite and half beidellite. The cristobalite-tridymite occurs as lepispheres 1 to 5 μ in diameter and as abundant smaller masses of less regular shape scattered through the smectite matrix. XRD patterns of the cristobalite-tridymite show opal-CT for the coarser fractions and opal-A for the finer, a consequence of both fine particle line broadening and increasing stacking disorder in the finer material. Less abundant minerals are quartz, feldspar, mica, and kaolinite. Unweathered beds contain calcite, sulfides, and organic matter.

INTRODUCTION

The Twiggs Clay is the argillaceous unit of a transgressive-regressive sequence that constitutes the Upper Eocene Jackson Group in east-central Georgia (Carver, 1980). It is a neritic unit with a maximum thickness of about 100 feet near Pikes Peak in Twiggs County, the type locality (Shearer, 1917; Cooke and Shearer, 1919), from which it thins in all directions. Typically, it is a gray to green, laminated to thin-bedded, sandy to hackly, diatomaceous clay. Updip in Bibb and Crawford Counties, it thins to less than 30 feet and is mainly sand. Downdip it thins and is increasingly calcareous; it merges toward the southeast into marl and shell beds and toward the southwest into the Ocala Limestone (Schmidt, 1977).

In the Pikes Peak area, the Twiggs Clay consists of two fuller's earth units separated by a bed of greenish sand (Schmidt, 1977, p. 8). The dry fuller's earth is mostly gray to light cream-colored and porous where it has been weathered. Its porosity is the result of removal, during weathering, of biogenic carbonate, some opal, organic matter, and fine sulfides. Where it is unweathered, it contains fine carbonaceous matter and fine sulfides and typically is gray-green to nearly black. Framboidal pyrite is abundant along partings and silty layers (Hurst, 1979). The clay is mostly thin-bedded, with fine sandy partings; a few beds are several feet thick. Numerous joints in the clay are coated with find secondary Mn-Fe-rich minerals, commonly in dendritic forms.

The mineralogy reported here applies to a suite of samples collected by one of the authors in 1956 from Diversey Corporation's fuller's earth mine in Twiggs County near Pikes Peak. The samples were taken from the face of a mine cut 20 feet high. They represent commercial fuller's earth, clays that had undergone sufficient weathering for destruction of fine sulfides, dissolution of biogenic carbonate, and partial mobilization of biogenic silica.

METHODS

Silty and sand layers were examined with an optical microscope. Fracture surfaces of clay layers coated with Au-Pd were examined with a Cambridge Mark II-A scanning electron microscope. For size fractionation the fuller's earth was dispersed by ultrasonification and fractionated by centrifugation. X-ray diffractograms of whole rock samples and various size fractions were made with a Philips X-ray diffractometer, CuK radiation, at a scanning rate of $1^o 2\Theta$/minute. Various treatments (glycolation, Li saturation plus heating plus glyceration, K saturation plus heating plus glyceration) were used in conjuction with XRD to identify the clay minerals. The proportions of smectite and cristobalite-tridymite were estimated from electron micrographs; the proportion of quartz was estimated by XRD (Hurst, 1956).

RESULTS

The silty parting between fuller's earth beds consists mainly of detrital quartz, feldspar, and mica with finer smectite (Fig. 1), the proportions varying. Some partings contain also kaolinite, cristobalite-tridymite, and calcite.

The principal minerals of the fuller's earth beds are smectite and cristobalite-tridymite. Less abundant minerals are quartz, mica, feldspar, kaolinite, and in unweathered beds calcite and sulfides, mainly pyrite. Estimated mineral percentages are 50 percent smectite, 30 percent cristobalite-tridymite, 15 percent quartz, and 5 percent other minerals. These estimated percentages are similar to those of Heron and others (1965) for Twiggs Clay from the General Reduction Company near Jeffersonville, Georgia, obtained by quantitative X-ray methods.

XRD patterns of different size fractions of the fuller's earth show the mineralogy gradually changing with particle size (Fig. 2). Smectite is a major component of all size fractions. Its proportion increases with decreasing particle size. In the less than 0.1 μ fraction, it is the only mineral detected by XRD, except a trace of kaolinite. Cristobalite-tridymite, formed by recrystallization of biogenic silica (Weaver and Wise, 1974), and detrital quartz and mica are more abundant in the coarser fractions. In the less than 0.1 μ fraction, cristobalite-tridymite is a minor component and no quartz or mica is detected.

Electron micrographs reveal that disaggregation of the fuller's earth prior to size fractionation was incomplete. Individual crystallites of smectite are all less than 0.1 μ, yet smectite is a major constituent of the coarser fractions. Individual crystallites of cristobalite-tridymite likewise are less than 0.1 μ in diameter, yet these minerals are most abundant in the 5 to 75 μ particle size range. Their abundance in the coarser fractions is due to their aggregation into lepispheres 0.6 to 5 μ in diameter and the attachment of the lepispheres to much coarser molds or tests of diatoms, sponge spicules, and other microfossils.

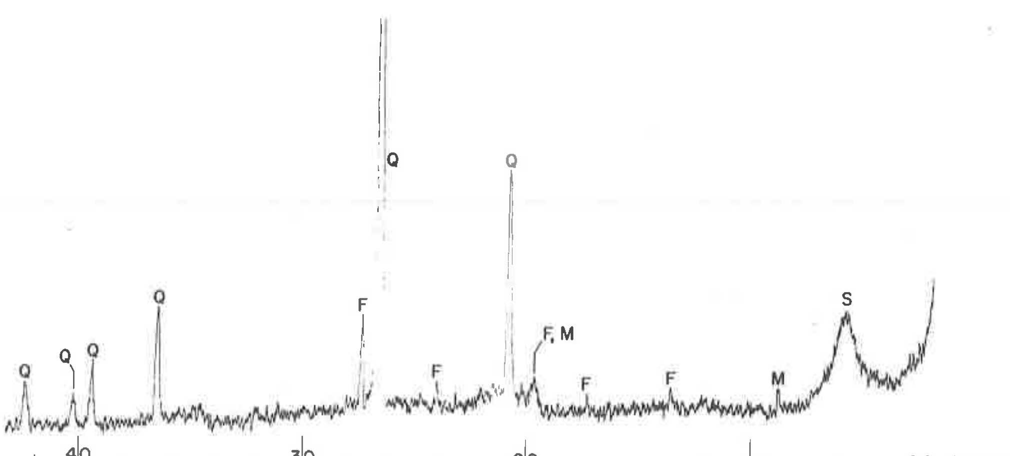


Figure 1. XRD pattern of a silty parting between fuller's earth beds. S = smectite, M = mica, F = feldspar, and Q = quartz.

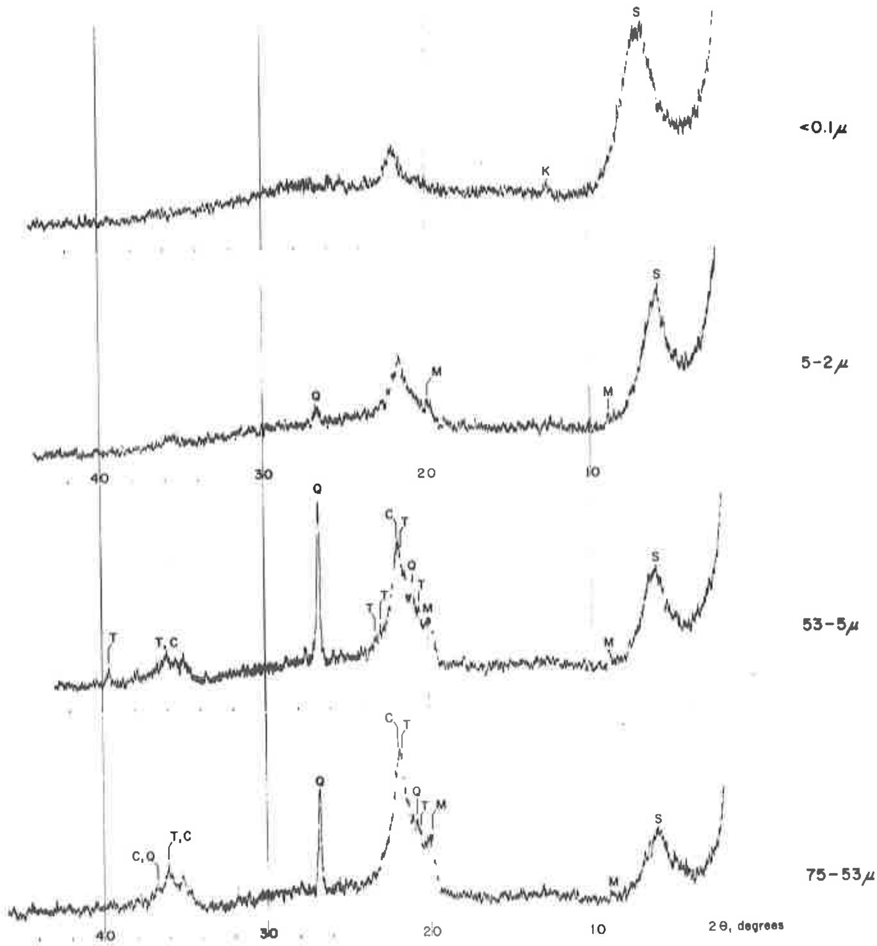


Figure 2. XRD patterns of size fractionated fuller's earth. S = smectite, M = mica, K = kaolinite, C = α-cristobalite, T = tridymite, and Q = quartz.

Smectite

XRD peaks of the major component, smectite, are conspicuously broadened due to fine crystallite size. The basal spacing of the untreated smectite is about 15 A. This spacing expands during glycolation to 18 A, and contracts at 600°C to 10.4 A (Fig. 3). The (060) spacing is 1.49, indicative of a dioctahedral mineral.

When a clay mineral whose layer charge imbalance is due mainly to octahedral charge deficiency, like montmorillonite, is Li-saturated and then heated, exchangeable lithium migrates from the interlayer positions to vacant octahedral sites. In this way the net negative charge of the octahedral sheet is compensated, after which the mineral behaves as a pyrophyllitelike structure which does not expand upon glyceration (Greene-Kelly, 1955).

When the Twiggs Clay is Li-saturated, heated overnight at 200°C, glycerated, and X-rayed, it shows two basal peaks of about the same intenstity, one about 19 A and the other at 9.6 A. This behavior indicates that only half the Twiggs Clay smectite is transformed by the Greene-Kelly treatment to a nonexpandable form, that is, only about half of the smectite is montmorillonite. The other half has appreciable charge deficiency in the tetrahedral sheet. Li saturation was done in a plastic centrifuge tube and the Li-saturated clay was heated in an inert crucible to prevent the spurious results reported by Brusewitz (1975).

K saturation of untreated clay followed by drying at 110°C for 3 hours, followed by glyceration, causes the clay to yield a broad basal XRD peak at 14.5 A. A minor peak at 10.5 A indicates that a little of the clay has contracted to a micalike structure not expandable by glyceration, the behavior of a mineral with a layer charge deficiency greater than that of montmorillonite, in this case either vermiculite or high-charge beidellite. K saturation and glyceration, without heating, yields a broad XRD peak interpreted as a combination of 14 and 18 A peaks. The 18 A peak obtained under these conditions (Fig. 3) connotes an expandable clay mineral with a c.e.c. less than 95 meq/100 gms of dry clay (Barshad, 1960), and the 14 A peak connotes an expandable clay mineral with a c.e.c. of 115 or greater.

Previously the Twiggs Clay smectite has been called montmorillonite (Brindley, 1957; Carver, 1972; Schmidt, 1977). The data presented above, which are more detailed than any previously reported, indicated that about half of the smectite here examined

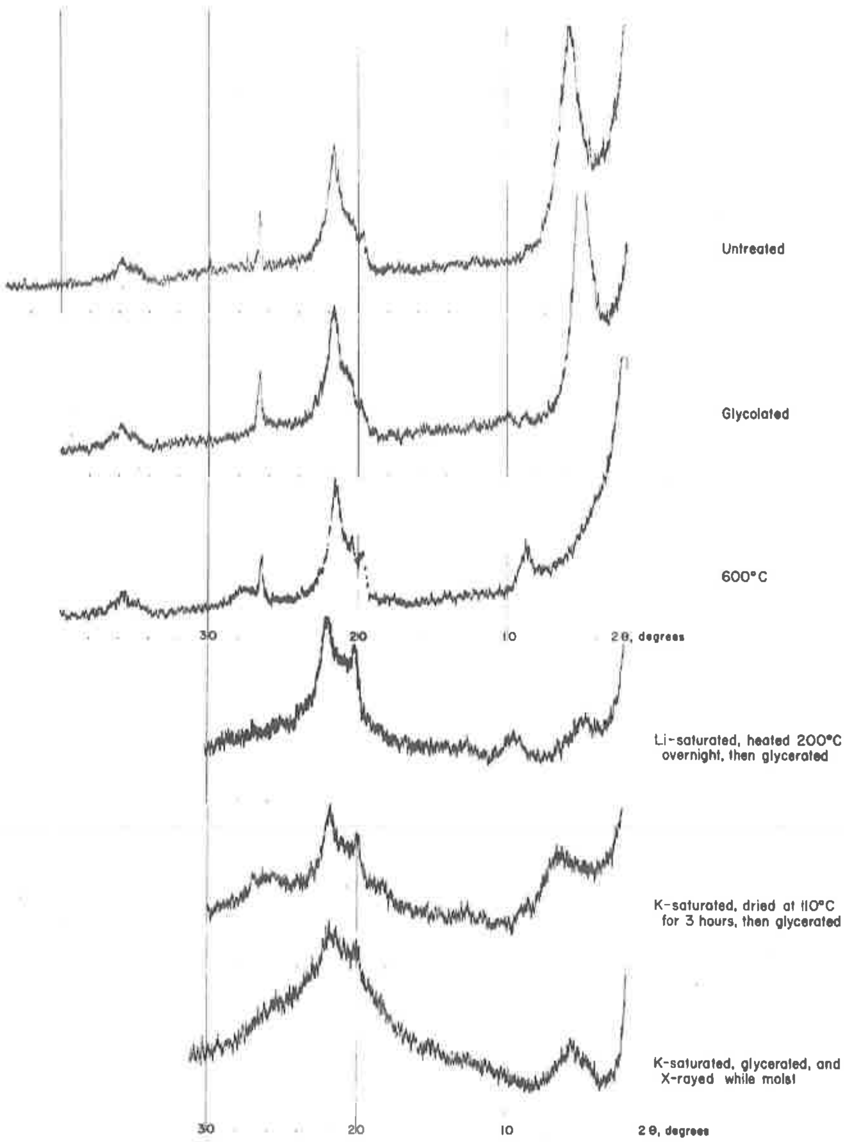


Figure 3. XRD patterns of the -0.1 μ fraction of the fuller's earth, untreated and after various treatments.

is montmorillonite, characterized by a moderate layer of charge deficiency originating mainly in the octahedral sheet, and that half of it is some other smectite characterized by a higher charge deficiency. This second smectite is dioctahedral, owes most of its charge to tetrahedral substitution, has a moderate iron content, as revealed by the light color of the oxidized sample, and a cation exchange capacity somewhat greater than that of montmorillonite. These are the characteristics of beidellite as defined by Greene-Kelly (1955) and Weir and Greene-Kelly (1962). The X-ray data indicate not a mixture of two distinct smectites, but a variety of smectites differing gradationally in composition and cation site occupancy and belonging to the montmorillonite-beidellite series, as discussed by Mering (1975). About half of the smectite can be categorized as montmorillonite and half as beidellite.

Cristobalite-tridymite

The second major component is cristobalite-tridymite, called opal-CT by Jones and Segnit (1971). Mainly it is in the form of lepispheres, 1 to 5 µ in diameter, concentrated in pores and concentrated on mold surfaces, where microfossils have dissolved out. Smaller lepispheres and even smaller masses of less regular shape are scattered through the smectite matrix.

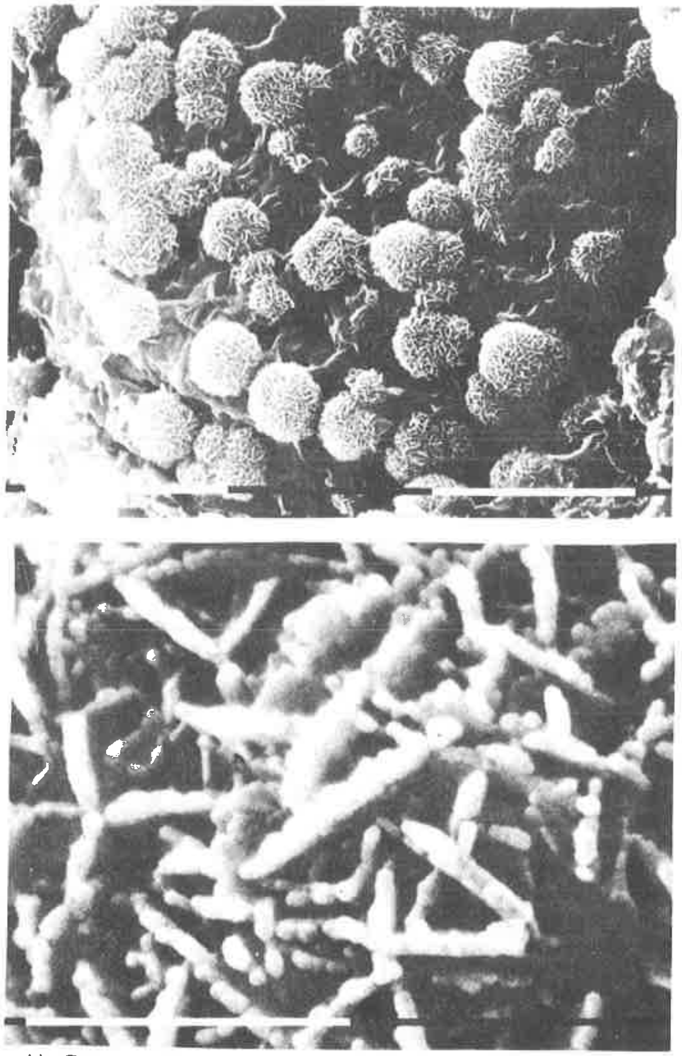
The lepispheres (Fig. 4A) have been described as microspheroidal clusters of bladed cristobalite (Wise and Hsu, 1971; Wise and Weaver, 1973). Magnification reveals that the "blades" actually are tabular aggregates of smaller equant masses 350 to 600 A in diameter (Fig. 4B). Whether these equant masses are themselves single crystals or aggregates of still smaller crystallites is not discernible from the SEM photos.

XRD patterns of silt-size fuller's earth (Fig. 5) show clear peaks of both α-cristobalite and tridymite (Table 1). While all of the peaks might be attributed to tridymite on the basis of position alone, the relative intensities of the peaks indicate both tridymite and cristobalite. In XRD patterns of finer fuller's earth, peaks are broadened and flattened to such a degree that only the more intense peaks common to both cristobalite and tridymite remain. The most intense reflection of both cristobalite and tridymite is from a layerlike linkage of SiO_4 tetrahedra about 4.1 A thick (Frondel, 1962, p. 288). In tridymite, two geometrically distinguishable sheets alternate, with the stacking sequence ABAB . . . while in cristobalite three geometrically distinguishable sheets alternate, their stacking sequence ABCABC . . . With increasing disorder in the stacking of these sheets, cristobalite and tridymite become indistinguishable by XRD. With total disorder in the stacking, only a low broad peak at about 4.1 A remains, called opal-A by Jones and Segnit (1971). The XRD distinction between opal-CT and opal-A does not necessarily relate to a mineralogical change, as a decrease in crystallinity, but may relate simply to a decrease in crystallite size. The very fine crystallites composing the "blades" of the lepispheres are fine enough to account for the few, low, broad peaks in the X-ray diffractograms of opal-CT. With decreasing size of crystallites, opal-CT peaks broaden and decrease in height until only the XRD patterns of opal-A remains. Florke and others (1976) have reported that lepispheres are groups of opal-CT blades intergrown according to two twin laws of tridymite. Figure 4B shows that the "blades" in some lepispheres, at least, are not simple crystallites of opal-CT but are tabular aggregates of even finer crystallites and that details of opal-CT crystallography are yet to be elucidated.

Table 1. d-spacings of cristobalite-tridymite peaks in XRD patterns of silt-sized Twiggs Clay fuller's earth.

d-spacing, A	Relative Intensity	
4.33	15	tridymite
4.10	100	α-cristobalite and tridymite
3.85	10	tridymite
2.50	40	α-cristobalite and tridymite
2.28	10	tridymite

That the cristobalite-tridymite is authigenic, the silica derived largely from diatoms and other siliceous microfossils--as previously suggested by Weaver and Wise (1979)--is apparent from their free growth morphology, the concentration of lepispheres



A

B

Figure 4. A) Concetration of coarser lepispheres of cristobalite-tridymite in the Twiggs Clay. This is a fracture surface of Twiggs Clay. The background material is mainly smectite. Bar equals 1 micron. B) Enlargment of one lepisphere, showing that the "blades" are tabular aggregates of nearly equant masses. Bar equals 1 micron.

on casts of microfossils--also found by Schmidt (1977)--and a rough inverse relationship between the abundance of lepispheres and the abundance of well-preserved siliceous microfossils, mainly diatom frustules and (less abundant) sponge spicules.

SUMMARY

At the type locality, the Twiggs Clay is a neritic, diatomaceous unit composed mainly of smectite and authigenic cristobalite-tridymite. Lesser minerals are detrital quartz, feldspar, mica, kaolinite, biogenic calcite, fine sulfides, and organic matter. Where the Twiggs Clay has been weathered, most of the calcite, sulfides, and organic matter have been destroyed, and what remains is a porous fuller's earth. Estimated mineral percentages are 50 percent smectite, 30 percent cristobalite-tridymite, 15 percent quartz, and 5 percent other minerals.

The Twiggs Clay smectite has variable composition and cation site occupancy. it

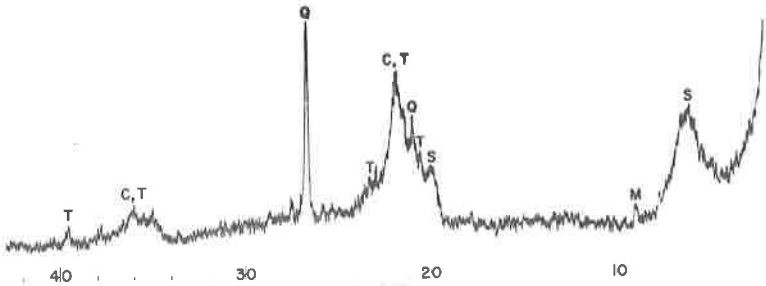


Figure 5. XRD pattern of silt-size fuller's earth showing α-cristobalite and tridymite peaks.

can be categorized as half montmorillonite and half beidellite. The cristobalite-tridymite occurs as lepispheres, mostly 1 to 5 μ in diameter, and as smaller masses of less regular shape. XRD patterns of the silt fraction show α-cristobalite-tridymite; XRD patterns of fine clay show only opal-A. Textures in SEM photos reveal that the cristobalite-tridymite is authigenic, the silica derived from diatoms and other siliceous microfossils.

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STRATIGRAPHY OF THE CHATTANOOGA SHALE, NORTHEASTERN TENNESSEE

By

Kenneth O. Hasson
Geography-Geology Department
East Tennessee State University
Johnson City, Tennessee 37614

ABSTRACT

The Chattanooga Shale of northeastern Tennessee is divided into three formal members, the nomenclature for which extends well-established Virginia terminology into the Tennessee outcrop. These members are: a lower black shale Millboro Member, a middle siltstone and shale Brallier Member, and an upper Big Stone Gap Member. The Brallier Member in Tennessee forms the southernmost Devonian marine deltaic lobe in the Appalachians, and is herein designated the Hawkins Lobe. Subdivision of the Chattanooga applies only to the Poor Valley outcrop belt.

The stratigraphic relationships within the Chattanooga are interpreted to result from a combination of sea-level fluctuation and differential subsidence within the area of Chattanooga deposition and the position of detrital clastic input into the upper Devonian sea. Local subsidence in conjunction with sea-level rise over the area began approximately in Early Frasnian, with deposition of up to 1305 feet (398 m) of black shale of the Millboro Member paraconformably over the Early Devonian Wildcat Valley Sandstone. A minor sea-level drop allowed coarser siltstones and silty shales of the Brallier Member to spill into the basin in the eastern (Poor Valley) outcrop belt, but these do not extent into the next northwestward outcrop belt. Westward spread of siltstones was prevented possibly by an intrabasinal high created by differential subsidence. A final sea-level rise produced the lower black shale part of the Big Stone Gap Member which upwards becomes gray, silty shale transitional to the regressive sediments of the Early Mississippian Grainger Formation.

INTRODUCTION

In the 90 years since being named by Hayes (1891), the Chattanooga Shale has received virtually no study in the folded outcrop areas of northeastern Tennessee, especially when compared to the massive efforts expended in the Highland Rim area during the uranium boom of the 1950's. Study of these strata is of particular importance in view of the active search for natural gas in rocks of this age and lithology, as exemplified by the U. S. Department of Energy's Eastern Gas Shales Project (referred to as DOE henceforth), and its recent (1980) completion of a Chattanooga test well in Grainger County, Tennessee. Also, a coring program to characterize the Chattanooga in eastern Tennessee has recently been completed by the DOE. In addition to its potential as a natural gas source, it is shale of this age and lithology which will be the primary feedstock for an eastern synfuel industy when the appropriate technology is developed.

The purposes of this paper are several: to describe in detail the stratigraphy of the Chattanooga shale in this area, to propose a nomenclature applicable where the formation can be subdivided, and to interpret the depositional history of the formation.

In northeastern Tennessee the Chattanooga Shale crops out in linear northeast-trending synclinal belts in Hawkins, Grainger, Hancock, and Claiborne Counties (Fig. 1). The longest and southeasternmost of these belts, approximately 60 miles (96 km) long, parallels Clinch Mountain in Hawkins and Grainger Counties and is labeled Poor Valley on topographic maps of the area. The northeastern part of this belt in Hawkins County is within the Greendale syncline, a structural feature which continues northeastward into Virginia.

Exposures occur also to the northwest of Poor Valley at Brushy Ridge and adjacent Newman Ridge, a syncline in Hancock County about 10 miles (16 km) northwest of Clinch Mountain (Fig. 1). A portion of the upper part of the Chattanooga is partly exposed on the northwest flank of the ridge, but the lower part is covered, occupying the valley of Blackwater Creek. The Chattanooga is about 500 feet (152 m) thick in these exposures and apparently of uniform lithology, but thins to about 326 feet (99 m) in cores #1 and #2 taken in Claiborne County (Roen and others, 1980). It is 605 feet (184 m) in core #3 (Fig. 1).

The Chattanooga ranges between 2000 and 600 feet (610-183 m) thickness in the Poor Valley outcrop belt, where it is divisible into three members: a lower Millboro Member, a middle Brallier Member, and an upper Big Stone Gap Member.

PREVIOUS WORK

Prior to the 1920's, study of the Chattanooga Shale (Hayes, 1891) in northeastern Tennesse consisted only of the large-scale geologic mapping done for the U.S. Geological Survey folios. Campbell (1894) used the term Chattanooga in the Estillville 30-minute quadrangle and recognized, but did not map separately, three parts: a lower black shale, a middle gray shale, and an upper black shale. Apparently Keith (1896, 1901) did not recognize these subdivisions while mapping the Morristown and Maynardville folios in Tennessee.

In a series of papers, Swartz (1924; 1926a, b; 1927; 1929a, b, c) examined various aspects of the Chattanooga, particularly its correlation and age relationships with more or less equivalent shales in southwestern Virginia. However, in 13 years of work in the area, I have never been able to approximate Swartz's sections.

Sanders (1952, 1963), in two unpublished manuscripts, recognized and named three informal members of the Chattanooga in the Greendale syncline area of Hawkins County, Tennessee. Hasson (1972) also recognized three members and tentatively adopted Sanders' nomenclature, but has since changed his preference, preferring to extend Virginia names with priority into Tennessee (Hasson, 1977). Published Chattanooga sections in the study area are: those of Swartz (1929), Dennison and Boucout (1974) who report a minimum of 2000 feet (610 m) of Chattanooga in cuts along Tennessee Route 70 at Clinch Mountain and describe the section in bed-by-bed detail, and Hasson (1973) who described in detail the Brallier and Big Stone Gap Members at two localities in the Poor Valley outcrop belt.

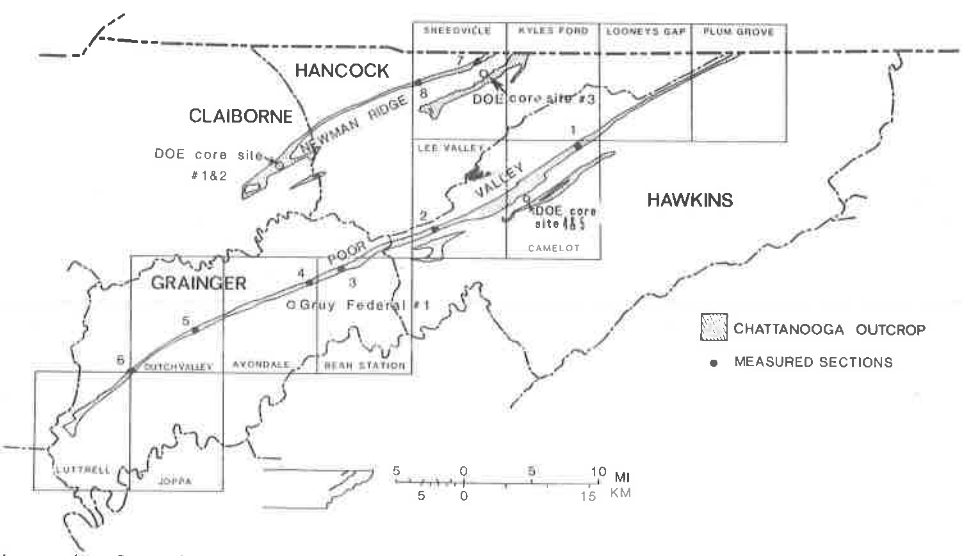

Figure 1. Location map of study area showing locations of measured sections, DOE cores, and DOE Chattanooga test well. Seven-and-one-half-minute quadrangles are also shown by name. Data point locations and Chattanooga outcrop trace are present-day base.

Very general outcrop descriptions and thickness estimates of the Chattanooga in this belt are given in the legends of the geologic maps of the Luttrell quadrangle (Swingle and others, 1967) and the adjacent Joppa quadrangle (Finlayson, 1965). Both authors recognized in an informal way the tripartite nature of the formation but attached no stratigraphic significance to it and did not attempt to map it in detail. Mixon and Harris (1971) also recognized the three-fold division and estimated a thickness of 850 feet (259 m) in the Poor Valley part of the Swan Island quadrangle. They also recognized the lack of divisibility in the Newman Ridge area (Harris and Mixon, 1970).

STRATIGRAPHY

The Chattanooga Shale in the study area consists primarily of thinly laminated grayish-black shale which weathers to large, thin, flat fragments. In the Poor Valley strike belt, thin to thickly bedded siltstone is locally intercalated with the shale.

The formation overlies paraconformably the Early Devonian Wildcat Valley Sandstone (Miller and others, 1964), most likely for the entire length of the Poor Valley outcrop belt. Miller and others (1977) record 3 meters (9 ft) of Wildcat Valley in the Luttrell quadrangle at the southwestern end of this outcrop belt and is also present at localities in the belt not noted by Miller and others (1977). Eighteen feet (5.5 m) of sandstone were encountered in the recent DOE test well (Gruy Federal #1 Grainger County) near Rutledge, Grainger County, Tennessee (Dean, 1980). If the Wildcat Valley is absent locally in the Poor Valley outcrop belt, the Chattanooga would overlie the Silurian Clinch Sandstone. Unfortunately, the basal contact of the Chattanooga is usually covered by Clinch Sandstone colluvium from higher up on the slope, and the continuous presence or absence of the Wildcat Valley is indeterminable. The cross section of Figure 2 assumes continuous Wildcat Valley. In the Newman Ridge area the Chattanooga overlies unconformably the Silurian Sneedville Limestone or the Early Devonian Wildcat Valley Sandstone. The Chattanooga is everwhere in the study area overlain conformably by the Early Mississippian Grainger Formation.

Poor Valley Belt

General Statement: Subdivision of the Chattanooga is possible only in the Poor Valley belt (Fig. 1). In much of this belt, three units are recognized: a lower Millboro Member of thinly to thickly laminated grayish-black shale; a middle Brallier Member of medium to very dark gray shale with interbedded siltstone; and an upper Big Stone Gap Member of fissile, grayish-black shale and gray, thickly laminated shale with thin siltstones interbedded near the top, grading upward into the Mississippian Grainger Formation. The Chattanooga-Grainger contact is placed at the base of a medium-to-thickly-bedded siltstone sequence (Hasson, 1972, 1973).

The members of the Chattanooga in this outcrop belt are defined solely on lithologic criteria and stratigraphic position. The proposed nomenclature utilizes well-established names extended into the study area from the adjacent region. The accompanying stratigraphic cross section (Fig. 2) illustrates the along-strike thickness and lithologic changes within the Chattanooga Shale southwestward between Tennessee Route 70 (Locality 1, Fig. 2) and the end of the outcrop belt in the Luttrell quadrangle. These relationships are discussed in detail in the following paragraphs.

In the Poor Valley Belt the Chattanooga Shale thins southwestward from a maximum of about 2000 feet (610 m) at Tennessee Highway 70 (1) (Dennison and Boucot, 1974), to approximately 600 feet (183 m) in the Luttrell quadrangle, as measured by plane table survey.

The southwestward-thinning along strike is accompanied by lithologic change. The gray silty shale of the Big Stone Gap Member becomes grayish-black shale; the thin siltstones and silty shales of the Brallier Member become massive siltstone and then again silty shale; the Millboro Member retains its black, fissile character.

Northeastward from the section at Tennesse Route 70 at Clinch Mountain (1), precise relationships are difficult to define because the Chattanooga is covered, with only rare partial exposures. Those few exposures at the position of the Brallier Member show greatly decreased silt content; distinct siltstone beds are lacking and the shale is only slightly silty. However, in the Virginia portion of this outcrop belt, a few miles northeast of the Tennessee-Virginia border in the Wallace quadrangle, Barlett and Webb (1971) record 100 ± feet (30 m) of Millboro Shale, 940 ± feet (287 m) of Brallier Formation, and 10 feet (3 m) of Big Stone Gap Shale, which here is entirely black, fissile shale. Between the Brallier and Big Stone Gap are 130 feet (40 m) of Chemung Formation marine sandstone, siltstone, and shale.

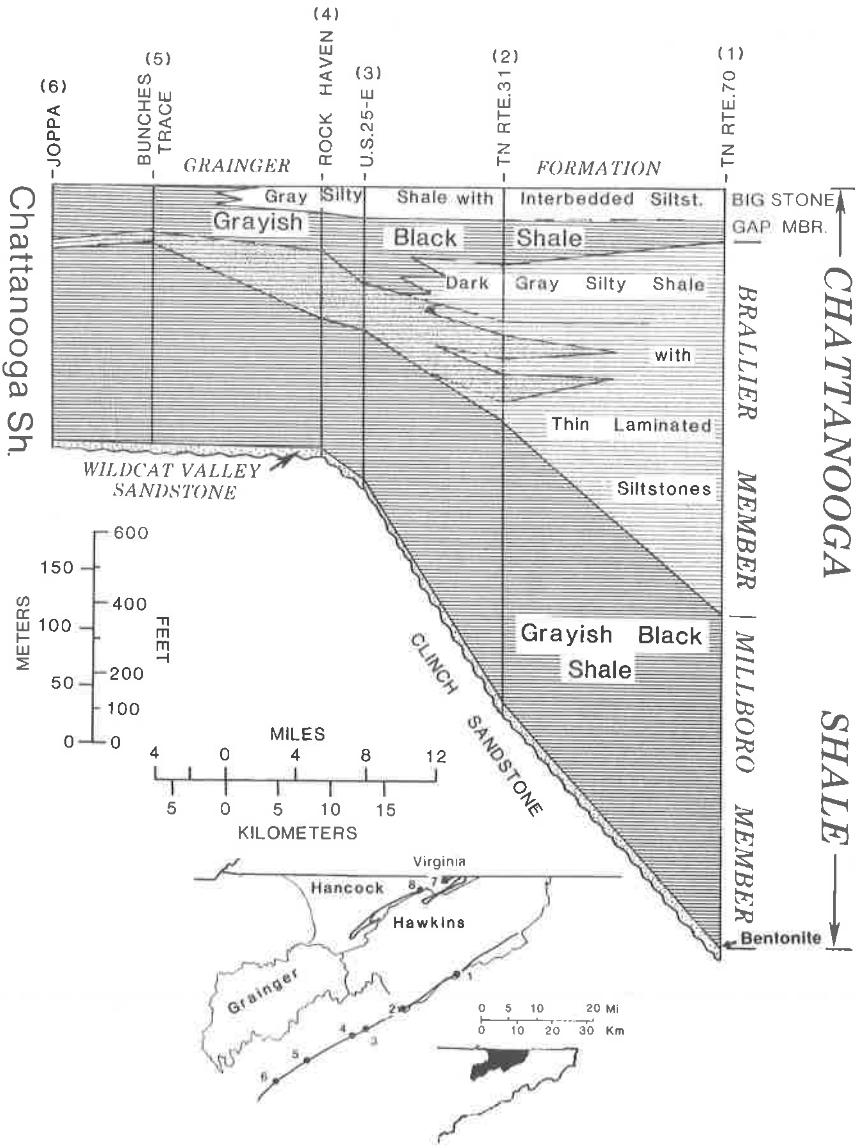

Figure 2. Stratigraphic cross section of Chattanooga Shale in the Poor Valley outcrop belt along southeast base of Clinch Mountain. Datum is base of the Grainger Formation (Hasson, 1972). The stippled pattern indicates the extent of massive siltstone beds in the Brallier Member, and is a section through the axis of the Hawkins Lobe. Section locations are shown on the outcrop trace restored to its predeformation position on a palinspastic base by Miller and others (1977). However, present-day geographic boundaries have not been shifted accordingly so that orientation accounts for the apparent shortening of the outcrop belt.

More detailed discussion of the individual members is given in the following sections, which are arranged in descending stratigraphic sequence, that is, in the order of drill penetration.

Big Stone Gap Member: Stose (1923) designated a sequence of black shale and siltstone below the Price Sandstone in southwest Virginia as the Big Stone Gap Shale. Since then, the unit has undergone a long and complicated history, which has been summarized in detail by Hasson (1972). Roen and others (1964) summarized the nomenclature applied to the black shale sequence in the vicinity of Big Stone Gap, Virginia, remeasured the sections of Stose (1923) and selected as typical the exposure on the southwest bank of the Powell River at Big Stone Gap, Virginia. They consider this unit as the Big Stone Gap Member of the Chattanooga Shale. In the type section the member is 242.6 feet (74 m) thick; the upper 151 feet (46 m) are shale and interbedded thin siltstones; the lower beds are gray to grayish-black shale (Roen and others, 1964, p. 346-347).

The most recent study of the type Big Stone Gap Shale is by Kepferle, Potter, and Pryor (1981). They remeasured the section and made a radioactivity log which they compared with that of a nearby well. On the basis of these logs they recognized the equivalents of the Sunbury, Bedford, and Cleveland Shales in the Big Stone Gap section; however, they also recognized the Big Stone Gap as a valid mapping unit.

The upper part of the Chattanooga Shale in the Poor Valley belt is assigned to the Big Stone Gap Member because of the similar stratigraphic position and lithologic sequence (Hasson, 1972, 1973). Here, the member consists of two parts: a lower fissile, grayish-black, platy weathering shale, and an upper gray shale with interbedded thin, nonlaminated siltstones. These shales and siltstones grade upward into the thickly bedded siltstones which define the base of the Grainger Formation. The lower black shale here included with the Big Stone Gap Member is apparently the Salt Lick Gap member (informal) mapped by Sanders (1952). That the Big Stone Gap Member in the study area has two, rather than three, parts is due probably to its proximity to sediment source and the availability of coarser sediment. Palinspasticly the closest exposure to the type section of the member in the Poor Valley belt is almost 40 miles southwest, or about 30 miles perpendicular to strike.

In the Poor Valley outcrop belt the lower black shale is probably the lithic equivalent of the Cleveland Member at the base of the type section, and the upper gray shale and siltstone part is equivalent to the Bedford and Sunbury. The Poor Valley area is far enough east for these units to have lost their lithic identity and to have merged to a more or less homogeneous mass.

The member is poorly exposed in this outcrop belt; the best, but not complete exposures occur in cuts along Tennessee Route 31 in Hawkins County and U. S. 25-E in Grainger County (locations 2 and 3 on Figs. 1 and 2).

The maximum thickness of the Big Stone Gap Member is 216 feet (66 m) as measured on Tennessee Route 31, with the lower black shale part being 119 feet (36 m) thick and the upper shale and siltstone 97 feet (30 m) thick. The upper gray shale portion grades laterally southwestward into grayish-black shale and thins to about 12 feet (4 m) in the Avondale quadrangle before it becomes unrecognizable. The thin siltstone beds, 0.05 to 0.12 foot (1.5 cm to 3 cm) which are abundant near the top of the gray shale in the Tennessee Route 31 and U. S. 25-E exposures, are absent to the southwest and the Grainger Formation overlies directly either gray or grayish-black shale. These relationships are shown in Figure 2.

Brallier Member: The Brallier Formation was named by Butts (1918) for exposures near Brallier Station, a railroad station about 5 miles (8 km) northeast of Everett, Bedford County, Pennsylvania. The Brallier consists of interbedded shale and silt-stone he shale is medium dark gray and weathers to olive or light olive gray chips. The siltstones are in sharply bounded beds 0.1 (.03 m) to 1 foot (.3 m) thick which weather to a characteristic rust color. Thickness of the Brallier Formation is on the order of 2000 feet (610 m) in the type area and northern Virginia. Bartlett and Webb (1971) report 940 feet (287 m) in the Wallace quadrangle, Washington County, Virginia.

Butts (1933, 1940) recognized and mapped the Brallier Formation the entire length of the valley in Virginia and noted that the belt southeast of Clinch Mountain in Virginia (Poor Valley Belt of this paper) continues into Tennessee some 40 miles (64 km), to the southwest end of Clinch Mountain (1940, p. 318).

However, because of the regional facies relationships, the Brallier in Tennessee is here considered more properly a member of the Chattanooga Shale, and, since the term Brallier has priority, should be used in place of other terms for this stratigraphic interval in this outcrop belt in Tennessee. The Brallier Member is not recognizable in outcrop belts or in cores to the northwest of Clinch Mountain, that is, on Newman or Brushy Ridges (Fig. 1).

The Brallier Member of the Chattanooga Shale underlies Poor Valley in the Tennessee portion of the Greendale syncline and is traceable southwestward along strike from Tennessee Highway 70 (Section 1, Fig. 1) in Hawkins County to the end of the outcrop belt in the Luttrell quadrangle (Grainger County). This interval is generally covered northeast of Highway 70 and the presence of the Brallier is suggested mainly by topography.

In the Highway 70 area the Brallier Member consists dominantly of dark to very dark gray, thinly to thickly laminated, silty shale which weathers to yellowish-gray plates and sheets. Thin (0.5 inch--1.2 cm--or less) dark gray, laminated siltstones are interbedded with the shale in certain intervals (Hasson, 1972; Dennison and Boucot, 1974). The laminae are light gray quartz in a darker micaceous silty matrix. The Brallier Member is about 1074 feet (327 m) thick at Highway 70 (Dennison and Boucot, 1974).

The member thins southwestward to about 150 feet (46 m) between Tennessee Route 31 (Section 2) and U. S. 25-E (Section 3) where the upper gray shale passes into the black shale facies of the Chattanooga (Hasson, 1972, and Fig. 2). Between these localities, the Brallier consists mostly of thickly bedded siltstones with maximum thickness centered on the Hawkins-Grainger County boundary (see Fig. 4). At Route 31 the massive siltstones are 180 feet (55 m) thick. The siltstones are both laminated and nonlaminated; the nonlaminated siltstones are thickly to very thickly bedded, with some individual beds 3 or more feet (0.9 m) thick. The massive siltstones produce a prominent ridge (Poor Valley Ridge), almost as high as Pine Mountain, which is held up by resistant beds of the overlying Grainger Formation.

The associated shales are generally thickly laminated, silty, dark gray and weather to a light olive gray color. The more thinly bedded siltstones generally are laminated, medium dark gray and weather to shaly plates. The gray shale overlying the siltstone is silty, thickly laminated, with a few thin interbedded siltstone beds up to 0.2 foot (6 cm) thick. The siltstone and the intervening strata between it and the overlying Grainger Formation thin progressively from Route 31 to the southwest (Fig. 2). The thickly bedded siltstone occurs at least 1 mile southwest of the end of Poor Valley Ridge on the Dutch Valley quadrangle, where it is 25 feet (8 m) thick. Pine Mountain ends abruptly and changes to a line of low hills which continues southwestward to the end of the outcrop belt in the Lutrell quadrangle. The topographic change marks the boundary between a dominant siltstone facies and a silty shale with minor siltstone facies.

From about the center of the Lee Valley quadrangle northeastward to the vicinity of the Tennessee Route 70 (Section 1, Figs. 1 and 2), the Brallier is characterized by a strike valley with distinct knobs. These knobs are held up by interbedded silty shale and thin siltstones.

Northeastward from Highway 70 to at least the Tennessee-Virginia state line, the height of the knobs diminishes according to decreasing silt content and loss of interbedded siltstone.

Millboro Member: The Millboro Shale (Butts, 1940, p. 309) was proposed as a mapping unit in Virginia with application to areas where the Marcellus-Naples portions of the Romney Formation (Tioga Bentonite to base of Brallier in current usage) could not be separated. The formation was named for Millboro Springs, Bath County, Virginia, where it is between 1325 feet and 1827 feet (404-557 m) thick, depending on the thickness in a covered interval at the base of the type section. The type section is described in detail by Hasson (1966) and Hasson and Dennison (1978).

Lithologically the Millboro is a grayish-black or very dark gray thinly laminated, platy to sheety-weathering shale, and is a mapping unit in Virginia from the type section south to the Tennessee border.

Its lithologic character and stratigraphic position indicate the Millboro to be contiguous with the lower black shale part of the Chattanooga in Tennessee, and it is proposed the term Millboro Member be applied to this interval. This formalizes the usage of Dennison and Boucout (1974), and provides an established term with priority for the rocks at this stratigraphic position.

The member consists principally of very dark gray to grayish-black, thinly to thickly laminated shale which weathers into yellowish or light gray plates and sheets. Locally it contains plant debris and is commonly very pyritic. Dip slopes show well developed jointing perpendicular to the bedding. The only published description of the Millboro Member in Tennessee is that of Dennison and Boucout (1974) who report a thickness of 1305 feet (398 m) at their Tennessee Route 70 section.

The Millboro Member overlies paraconformably the Early Devonian Wildcat Valley Sandstone. The contact and lower part of the member are particularly well exposed in the cut on Tennessee Route 70 over Clinch Mountain described by Dennison and Boucout (1974), and in recent construction of U. S. Route 25-E over Clinch Mountain.

Dennison and Boucout (1974) also describe the Early Devonian Tioga Bentonite as being present through some 22 to 27 feet (6.7 to 8.2 m) of strata above the Wildcat Valley Sandstone. However, conodont studies by Anita Harris of the U. S. Geological Survey on samples submitted by Roy C. Kepferle in the course of his studies of the Devonian black shales of the Appalachian basin, indicate these beds to be earliest Late Devonian and possibly represent the Center Hill Bentonite (Anita Harris, personal communication, 1977). Roen (1980), in an unpublished report cited in Kepferle and Roen (1981), equates these ash beds to the Belpre Bentonite Bed of Collins (1979). This (or possibly another) bentonite is present in the DOE Grainger County #1 well 86 feet (26 m) above the Wildcat Valley Sandstone (Claude Dean, oral communication, 1979). This bentonite should provide a marker bed in any future drilling in this area. The Millboro Member is nowhere completely exposed because it weathers easily and normally underlies Poor Valley or is covered with colluvium. Thickness measurements are only best estimates because the unit is structurally incompetent.

Newman Ridge Area

The Chattanooga Shale in the Newman Ridge area overlies unconformably the Silurian Sneedville Limestone or locally the Wildcat Valley Sandstone in a belt on the northwest side of the ridge (Murray, 1970). The Chattanooga underlies the valley of Blackwater Creek and is mostly covered with only the upper 50 to 100 feet (15 to 30 m) of the formation exposed in cuts on the two roads across Newman Ridge. Where exposed, the Chattanooga is very dark gray to grayish-black, thinly to thickly laminated pyritic shale which weathers to plates or sheets. Thickness is about 500 feet (152 m) on the northwest face of Newman Ridge as measured by plane table survey, but this figure is really a best estimate because of cover and structural incompetence.

Roen and others (1980) report 326 feet (99 m) of Chattanooga in DOE cores #1 and #2 in the Howard Quarter quadrangle, Claiborne County, Tennessee, and 605 feet (144 m) in core #3 (Fig. 1). They describe the Chattanooga as dominantly (75%) brownish-black or grayish-black shale, the remainder being gray to greenish gray clay shale and siltstone. The siltstones are 0.02 foot (0.6 cm) to less than 1 millimeter thick and are not abundant. The shale is laminated to very thinly bedded with some thick to massive beds.

The subdivisions of the Chattanooga recognizable in the Poor Valley belt are not apparent in the Newman Ridge area, although some knobby topography similar to that produced by the Brallier Member occurs near the Virginia-Tennessee border in the Sneedville quadrangle. This may represent an incursion of Brallier lithology into this area, but it is too feebly developed and does not extend sufficiently far south to divide the mass of black shale into members recognized in the Poor Valley belt.

DISCUSSION

In the Poor Valley outcrop belt the Chattanooga shale rests paraconformably on the Early Devonian Wildcat Valley Sandstone. This sandstone includes beds ranging in age from Helderberg to Onesquethaw, although at no place is the sequence complete. At Tennessee Route 70, Helderberg strata are absent and a minor unconformity separates Oriskany and Onesquethaw-age beds (Dennison and Boucout, 1974). Obviously sea level was quite low and sediment removal and redistribution easily accomplished.

Following deposition of the Onesquethaw-age sandstones, the sea apparently withdrew from this part of the southern Appalachians, with the shoreline stabilizing in the general vicinity of Hayter Gap, Washington County, Virginia (Fig. 3).

Evidence for the presence of the mid-Devonian sea in the Hayter Gap area of southwestern Virginia is provided by House (1962), who describes an ammonoid fauna from the lower Millboro Shale there. The fauna is correlative with the upper part of the ***Cabrieroceras crispiforme*** Zone of Devon (mid-Givetian). That this may have been a marginal marine area and an unconformity may exist within the Millboro Shale is suggested by the fact that the mid-Givetian fauna is overlain almost immediately by a ***Probeloceras*** fauna comparable to the Frasnian Cashaqua of New York (House, personal communication, 1976). Later, he (House, 1978, p. 12) remarked that, "***P. lutheri*** follows a short distance above the ***Sobolewia virginiana*** Goniatite bed suggesting either a stratal gap, or substantial reduction."

Further evidence for withdrawal of the sea from Tennessee and the adjacent southwest Virginia area is provided by the previously cited work of Anita Harris who identified earliest Late Devonian (Frasnian) conodonts in samples from the base of the Chattanooga and between the "Tioga" bentonite layers in Dennison and Boucout's (1974) described section (Anita Harris, personal communication, 1977).

Harris' work demonstrates the absence of Middle Devonian strata at this locality and precludes the bentonite described as being Tioga. A reasonable possibility is that the "Tioga" is the younger Center Hill Bentonite or possibly even the Belpre Bentonite.

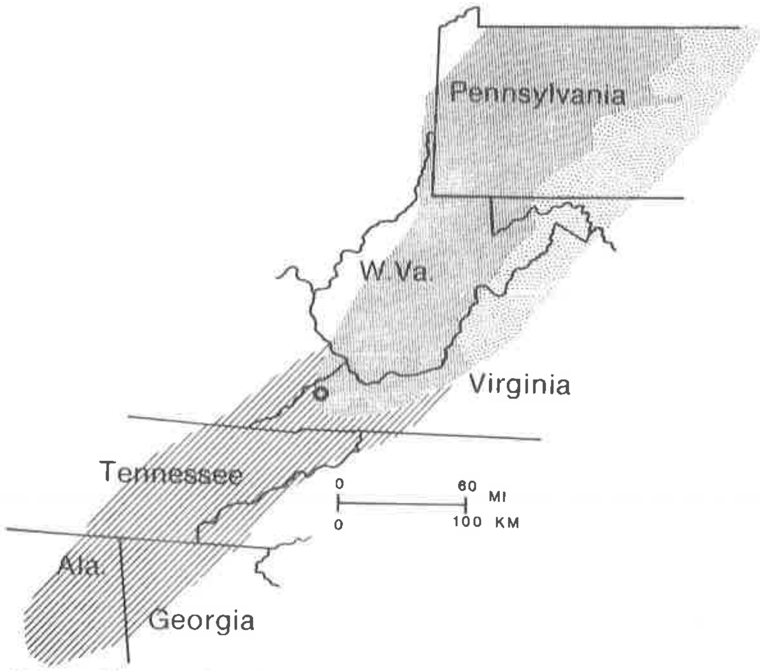

Figure 3. Schematic map showing extent of inferred withdrawal of the sea following deposition of the Wildcat Valley Sandstone. Area uncovered by withdrawal is indicated by diagonal stripes. The vertical wavy lines indicate persistent sea. The black circle is Hayter Gap, Virginia. The sand pattern indicates Middle and Late Devonian deltaic lobes, which were areas of continual deposition while the southern Appalachian basin was subaerially exposed.

The contact between the Early Devonian Wildcat Valley and the Late Devonian Chattanooga is paraconformable, there being no obvious evidence of erosional unconformity. The Early Devonian sea withdrew from this evidently flat surface with no apparent tectonic activity, and remained restricted to southwestern Virginia during the Middle Devonian.

Miller and others (1977) have suggested leaching of the calcite-cemented-wildcat Valley Sandstone during the 15 or so million years of the Middle Devonian to produce a porous and permeable upper layer. Subsequent calcite cement is derived from fresh water and some of the original primary porosity may be preserved, which would make the Wilcat Valley Sandstone an excellent potential reservoir rock.

Recent construction on U. S. 25-E in Grainger County, Tennessee, near Bean Station, has exposed the Wildcat Valley-Chattanooga contact. In this exposure the upper half-foot of Wildcat Valley Sandstone is pyrite-cemented sandstone sharply overlain by black, pyritic shale. The pyrite cement would indicate the existence of primary porosity from Middle Devonian weathering which allowed iron-bearing waters to percolate downward on the weathered surface as suggested by Miller and others (1977).

Microscopic examination of polished sections reveals shale stringers penetrating downward into the spaces between the quartz grains, which is also suggestive of inundation of a friable sand surface. These relationships are shown in the photomicrographs in Figure 4 (A, B).

The pyritic sandstone and shale at this locality also lends support to the following scenario. Near or at the end of the Middle Devonian, the sea, restricted to southwestern Virginia until this time, began to expand. The Taghanic onlap, a continent-wide eustatic sea-level rise, produced an eastward shift of black shale facies in the central Appalachians (Hasson and Dennison, 1973, 1978), with concomitant expansion of anoxic waters to the south and west. This resulted in the paraconformable deposition of pyritic black shale of the Millboro Member on the near-base-level surface of Wildcat Valley Sandstone in Tennessee and southwest Virginia.

In northeastern Tennessee deposition of the black shale was most likely accompanied by local differential subsidence. Subsidence is necessary to accommodate from between 2000 and 600 feet (160 to 183 m) of shale on what was a surface at or very near sea level, as suggested by me previously (Hasson, 1974, 1978, 1980). At least 1600 feet of Chattanooga were drilled in DOE cores #4 and #5 a few miles southeast of Tennessee Route 70; this is a much thicker section than that at Tennessee Route 31. The increased thickness is attributed to depositional causes (Kepferle and Roen, 1981). The necessity for Late Devonian subsidence in the eastTennessee area was also concluded by Lundegard and others (1980) to explain discrepancies in their water-depth calculations for the Chattanooga. The thick accumulation of black shale in the Greendale and Newman synclines requires deepening of the basin at this time. Elsewhere, even at the base Chilhowee Mountain to the southeast of Knoxville, which would be much closer to a possible eastern source (paleocurrents were almost east-west during black-shale deposition; Jones and Dennison, 1970), the entire Chattanooga is only 25 feet (8 m) or less thick (Neuman and Nelson, 1965).

The introduction of Brallier Member turbidites and silty shale into the Chattanooga produced the southernmost Devonian deltaic clastic lobe in the Appalachians (Hasson, 1978). The siltstones in this lobe probably result from a combination of regional tectonism to the east and a minor lowering of sea level which allowed the silts to be carried into the deeper part of the basin. The Brallier lithology siltstones also constitute the southernmost record of the earlier phases of the Acadian Orogeny and extend the Catskill clastic wedge into Tennessee. Lundegard and others (1980, p. 57) also intrepreted the Brallier Member "as a turbidite lobe deposited basinward of a westward-prograding delta."

Previously, in a 1974 paper in Atlanta, I referred informally to this marine deltaic lobe as the Hawkins Lobe; I would now formally designate the Hawkins Lobe for its exposure and development in Hawkins County, Tennessee. The location of the Hawkins Lobe in Tennessee is shown on Figure 5.

The silty shales and siltstones of the Hawkins Lobe are succeeded directly by black shales of the basal part of the Big Stone Gap Member. These shales become gray and silty with increasing numbers of thin, discrete siltstones, and are transitional to

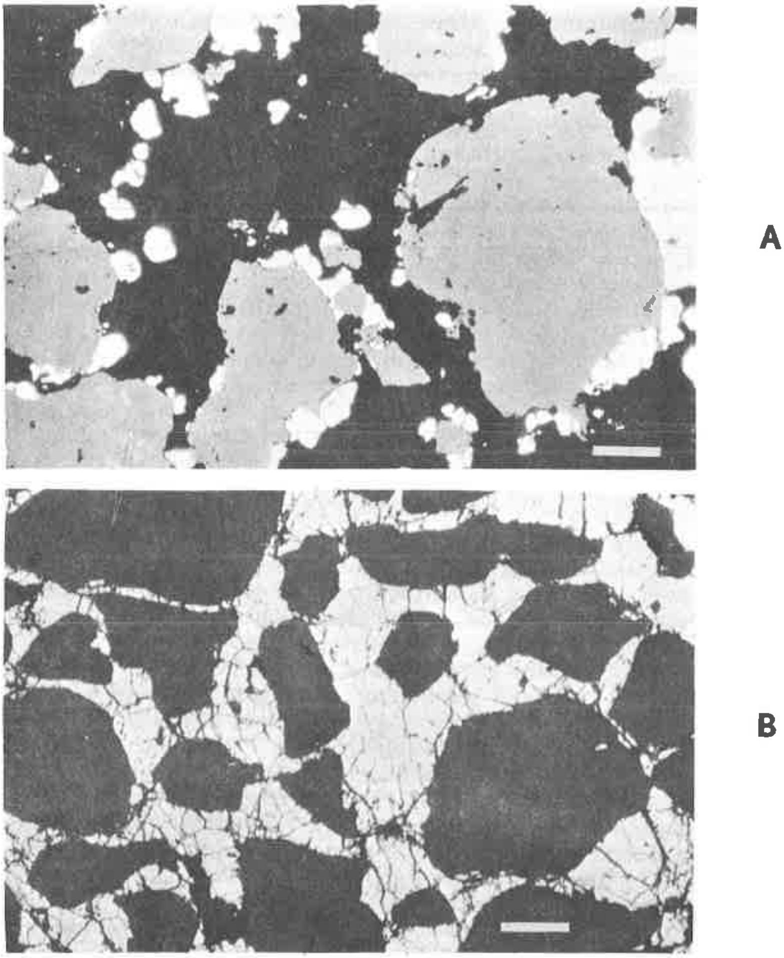

Figure 4. Photomicrographs of uppermost Wildcat Valley Sandstone in sample from exposure on U. S. 25-E., Grainger County, Tennessee. A. Area near basal Chattanooga Shale-Wildcat Valley Sandstone contact. This photo illustrates shale infiltration into the uppermost part of the Wildcat Valley Sandstone, suggesting surface friability at the time of Chattanooga inundation. Bright white grains are pyrite; gray is quartz; and black is black, pyritic shale. Long dimension is 1150 microns; the bar is 100 microns long. Reflected light, 40 X. B. Pyrite-cemented sandstone a few millimeters below A. The two photos together suggest that pyrite was precipitated early and prior to the influx of Chattanooga muds. In some hand specimens quartz grains float a millimeter or two above the shale-sandstone contact. Gray is quartz, light color is pyrite, and black is shale. Long dimension is 1150 microns; bar is 100 microns long. Reflected light, 40 X.

overlying Mississippian Grainger Formation.

Deepening of the water over the Hawkins Lobe to produce an environment suitable for black shale deposition is indicative of a sea-level rise at this time which effectively shifted the sediment source eastward, cutting off supply of coarser clastics. This Late Devonian rise appears to be widespread, represented not only by the lower part of the Big Stone Gap Shale but also by such units as the Exshaw, Riddlesburg, New Albany, and Ohio Shales. Effects of such a sea-level rise are more evident on marginal areas; regional relationships of transitional Devonian-Mississippian strata along the cratonic margin have been described by Gutschick and Moreman (1967), and in most cases the boundary lithologies involve shale at this position.

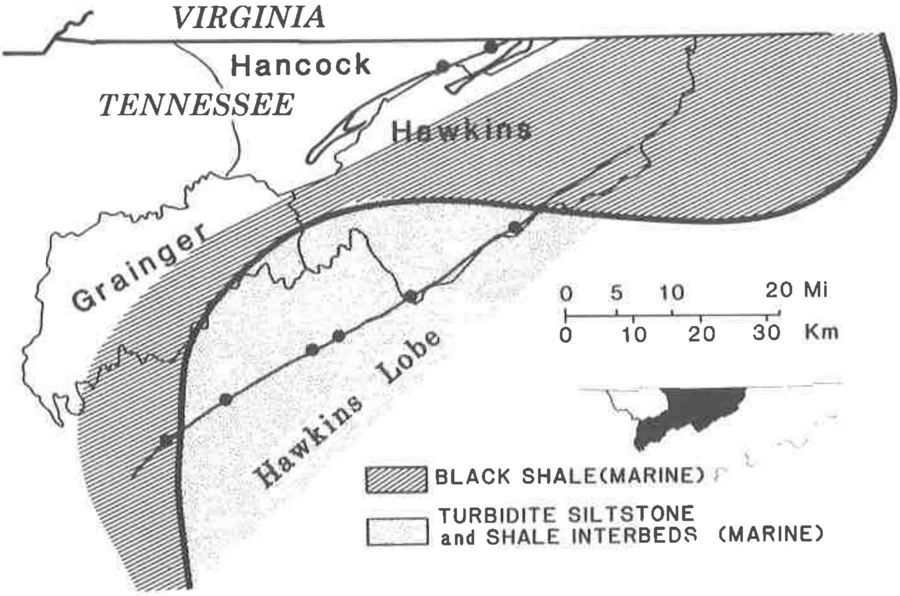

Figure 5. Outline of Hawkins Lobe and Scott Bay on palinspastic base but present-day geographic boundaries. The outcrop trace location is from map by Miller and others, 1977. The outcrop trace appears shortened because geographic boundaries have not been shifted; the northeastern end of the outcrop trace is at the present Tennessee-Virginia border.

An episode of Devonian faulting in the southern Appalachians has been suggested recently by Roeder (1975, 1978). The field relations as described in the preceeding paragraphs indicate no apparent sedimentologic break in deposition of Chattanooga sediments, such as those that would suggest a major tectonic disturbance. To the contrary, the field evidence suggests relative quiescence during the Devonian.

CONCLUSIONS

The Chattanooga Shale of northeastern Tennessee is divisable into three members in the Poor Valley outcrop belt. In ascending order these are the Millboro, Brallier, and Big Stone Gap Members.

The Devonian strata exposed in the study area reflect a somewhat complicated depositional history involving three transgressive phases and three regressive phases superimposed on a differentially subsiding area. This series of events is summarized in the sea-level curve of Figure 6.

The initial transgression produced shallow flooding over the Wallbridge surface (Wildcat Valley Sandstone) followed by a regression which lasted through all or most of the Middle Devonian, producing a weathered surface on the Wildcat Valley Sandstone. The next transgression occurs at about the mid-Late Devonian boundary and is the southern extension of the Taghanic onlap. This sea-level rise, which produced the black shale of the Millboro Member, was coupled with differential subsidence of the near-base-level surface. This combination of events is necessary to accommodate the thick accumulation of shale.

The Brallier Member silty shales and turbidite siltstones represent a minor regression produced by a combination of a westward-prograding delta and a minor sea-level lowering. Differential subsidence of the basin was sufficient to produce a clastic trap, because the silts and silty shales of the Brallier Member do not extend northwestward into the next outcrop belt in Tennessee.

The uppermost part of the Brallier Member in the Poor Valley outcrop belt corresponds in stratigraphic position to the Three Lick Bed (Provo and others, 1977) in the Big Stone Gap area. The Three Lick Bed may be in reality the basinward extension

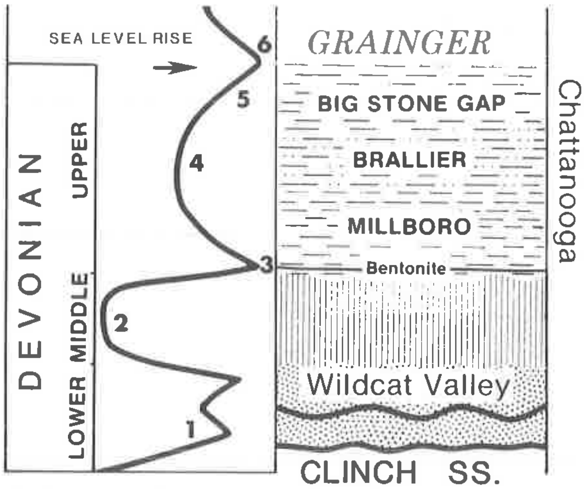

Figure 6. Combined stratigraphic section and sea-level curve for area described in this paper. The position of the unconformity in the Wildcat Valley is schematic. The column illustrates the following sequence of sea-level fluctuations. (1) Rise over pre-Devonian unconformity; (2) Middle Devonian withdrawal; (3) Early Frasnian rise (Taghanic onlap); (4) minor lowering in Brallier time; (5) rise to produce black shales of lower Big Stone Gap; (6) gradual lowering of sea level to produce regressive deltaic sequence of Early Mississippian Grainger Formation. The sea-level fluctuations must be coupled with local subsidence to accommodate the thick succession of Chattanooga Shale on the Wildcat Valley surface.

of the uppermost Brallier. However, Provo and others do not recognize the Three Lick Bed in the easternmost outcrops. In this area gray shale and silt-stone are the dominant lithologies, replacing the Three Lick bundle of three beds of greenish-gray shale separated by black shale.

Black shale of the lower Big Stone Gap Member represents the next transgression,which is coupled with the regressive phase of the Early Mississippian Grainger Formation. The Grainger is a basing-filling deltaic formation and indicates a progressively diminishing rate of differential subsidence from Late Devonian to Early Mississippian. A similar coupling of differential subsidence and transgressive-regressive phases in the Devonian-Carboniferous of Ireland has been described recently by McCarthy and Gardiner (1980).

The stratigraphic succession observed in the field can be very satisfactorily accounted for by facies change and local subsidence. Devonian faulting, as suggested by Roeder (1975, 1978) is not supported by any field evidence.

ACKNOWLEDGMENTS

Field work for this study was initiated in 1968 and has continued intermittenly to the present: some of the data are included in a dissertation by Hasson (1972), but much additional work has led to a greatly increased understanding of the stratigraphy of the Chattanooga Shale in northeastern Tennessee.

During this work I have been aided greatly by students, two of whom, James Barrett and Steve Bagby, were of particular help as field assistants. Walter Wheeler critiqued the manuscript. Thanks also are due Steve Haase, Oak Ridge National Laboratory, for making the photomicrographs. Initial field work was supported by a grant from East Tennessee State University with subsequent work underwritten by the author.

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ERRATUM FOR VOLUME 23, NUMBER 1

PLUTONIC EVENTS IN THE PIEDMONT OF VIRGINIA

By

Douglas G. Mose and M. S. Nagel

Insert the following paragraph at the bottom of page 26:

The Quantico slate has been correlated with the Arvonia slate in the Arvonia and Columbia synclines in central Virginia and Late Ordovician fossils have been found in the Arvonia slate (Tillman, 1970; Higgins, 1972; Higgins and others, 1977; Brown, 1979; Pavlides, 1980). The Arvonia slate is mainly a graphitic slate which unconformably overlies the Chopawamsic Formation, and the Chopawamsic Formation along with the underlying Evington Group rocks are more intensely deformed than the slate and show at least one stage of folding not present in the Arvonia slate (Brown, 1970a, 1979).