# SOUTHEASTERN

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# STATISTICAL STUDY OF ZIRCON POPULATIONS FROM

# IGNEOUS AND METAMORPHIC ROCKS AS A METHOD

# OF DETERMINING MIXED POPULATIONS

By

James F. Conley

and

Oliver M. Fordham, Jr.
Virginia Division of Mineral Resources
Charlottesville, Virginia 22903

## ABSTRACT

A statistical study was made of zircons from rocks in the Martinsville West quadrangle, in the southwestern Piedmont of Virginia. Computer programs were written for statistical comparison of zircon populations using computer contouring of length-breadth data and the reduced major axis method.

Results of the study show that the igneous and metasedimentary rocks of this area contain multiple populations of zircons. Extremely poor correlations were obtained from zircon populations from the igneous rocks (even from the same unit). Present work indicates that the reduced major axis method, considered the best statistical tool for comparing zircon populations, should not be used alone as it assumes that the data follows a normal distribution. Serious misinterpretations could result when dealing with igneous rocks containing multiple populations of metadetrital and igneous zircons. Lack of correlation among zircons from igneous rocks might indicate multiple populations from contamination rather than lack of correlation between the igneous populations from the same magma. Thus, lack of correlation among zircon populations could be used as a first indicator of multiple populations.

## INTRODUCTION

The Martinsville West 7.5-minute quadrangle, located in the southwestern Piedmont of Virginia, about 20 miles east of the Blue Ridge (Figure 1), was mapped by Conley and Toewe (1968). The zircons studied were obtained from rocks within the Smith River allochthon, a synformal structure that is situated between the Blue Ridge and Sauratown Mountains anticlinoria and is surrounded by, and structurally
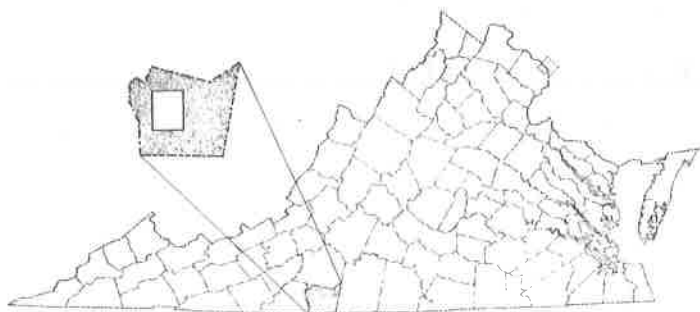
Figure 1.  Index map showing locations of study area (Martinsville West quadrangle, Virginia).

overlies rocks of Grenville and younger Precambrian age. The oldest rocks in the allochthon are of metasedimentary origin. These rocks were intruded by a multi-injection igneous complex, which, in the vicinity of its contact, has converted the metasedimentary rock into gneissic hornfels (Conley and Henika, 1973).

Poldervaart (1956) and Larsen and Poldervaart (1957) have found that zircons, especially in felsic rocks, crystallize over a short range and before the main constituent minerals form. Therefore, zircons crystallizing in a particular magma should have morphological characteristics unique to that magma and zircons derived from that magma should be statistically comparable and should contrast with those derived from other sources. Contrary to this, Silver and Deutsch (1963, p. 756) have found that zircons begin to crystallize early in a melt and grow over much of the crystallization history of the parent rock. If the latter is correct, it might indicate that variations should occur in separate sample populations derived from the same magma. This might explain the phantom overgrowths, distortion of side prisms, and the elongate euhedral nuclei observed in zircon populations during this study. Additionally, contamination by detrital grains incorporated in the magma from the country rock could also cause variations in populations.

## Acknowledgments

## OBJECTIVES

The objectives of this study were: 1) to determine if relict detrital-metamorphic zircons were present in the plutonic rocks which make up the igneous complex, 2) to see if rocks of the same lithologic type from different parts of the igneous complex could be correlated with each other and 3) to determine if an igneous or detrital-metamorphic genesis could be indicated for zircon populations collected from migmatitic country rock intruded by the igneous complex. Computer programs were devised for mathematically comparing various zircon populations.

Extremely poor correlations were obtained from zircon populations taken from rocks of the igneous complex (even populations from rocks of the same mapping unit) using the reduced major axis method for statistical comparison (Tables 1, 2, and 3). These findings seem to be in contrast to those previously published (Larsen and Poldervaart, 1957, Alper and Poldervaart, 1957, and Taubernack, 1957). For this reason further research was initiated into the cause of this lack of agreement and the findings of this work is here presented.

## SAMPLE COLLECTION AND ZIRCON CONCENTRATION

Ten samples were collected for this study (Figure 2): five from granite (R-3380*, R-3383, R-3386, R-3387, and R-3389); one from gabbro (R-3382); two from biotite paragneiss, a layered biotite paragneiss (R-3381) and a migmatitic biotite paragneiss (R-3385); two samples of pegmatite, a coarse-grained inequigranular pegmatite cutting amphibolite (R-3384) and an equigranular alaskitic pegmatite cutting the gabbro (R-3388).

Almost pure zircon concentrates an be made using various procedures described by Larsen and F rvaart (1957), Drummond (1962), and Neuerburg (1975). Zircon concentrates for each sample are mounted on glass slides with Lakeside 70 thermoplastic cement. Hedberg and Greenwood (1970, p. 2) recommend Hydrax as a mounting media because it has a refractive index of 1.823 and limits the refractive shadow of zircon grain boundaries.

## DATA COLLECTION

After each zircon population is mounted, equal-spaced traverses are made across each slide using a mechanical stage equipped for

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*Numbers preceded by "R" are Virginia Division of Mineral Resources repository numbers for rock samples.

EXPLANATION

GRANITE

GABBRO

SCHIST

AMPHIBOLITE

BIOTITE PARAGNEISS
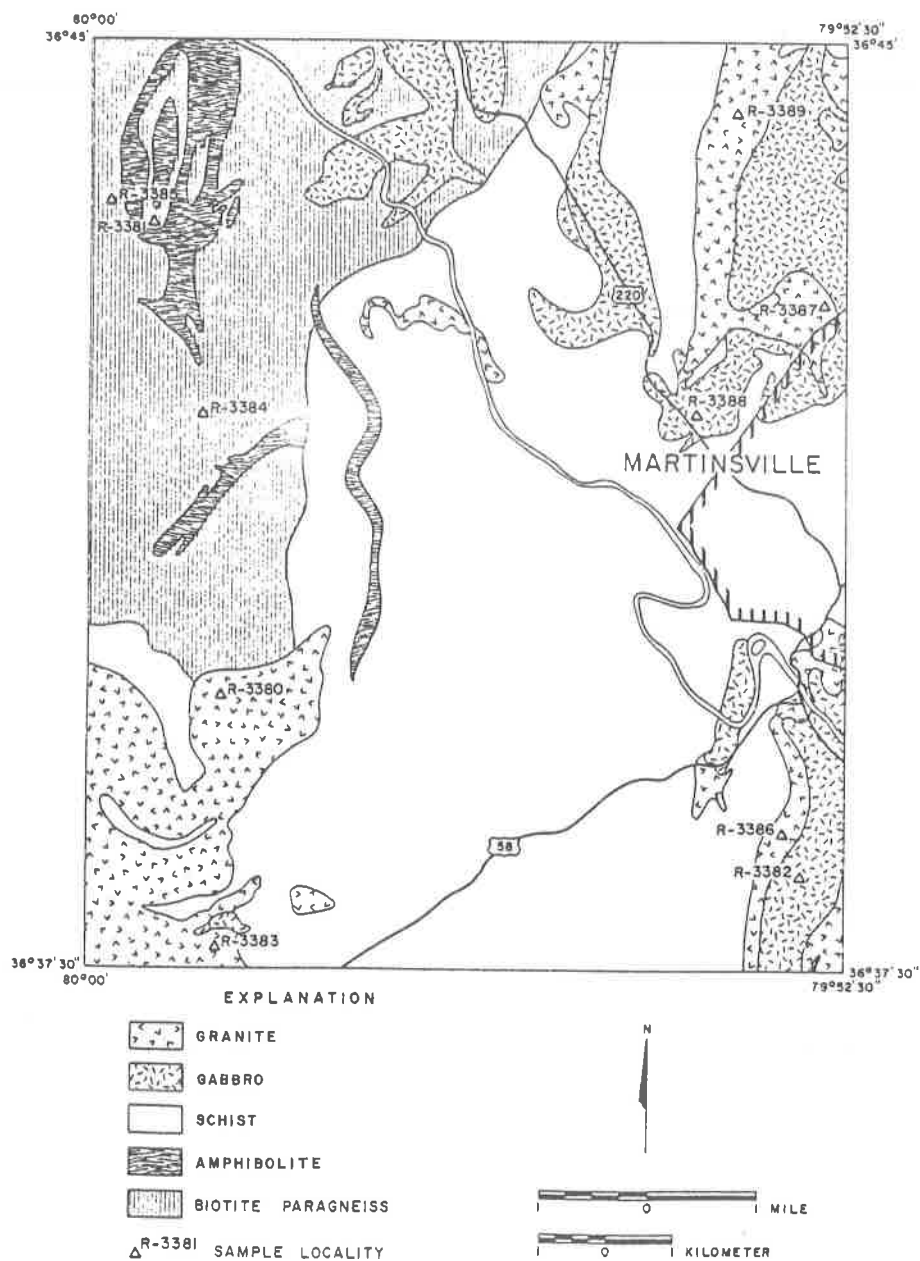
△R-3381  SAMPLE LOCALITY

Figure 2.  Geologic map of Martinsville West quadrangle, Virginia showing sample localities.

point counting. As the mechanical stage is advanced, each doubly ter-
minated zircon crystal that falls under the cross hairs of the micro-
scope, is measured for length and breadth using an ocular micrometer
at a magnification of 400x until 200 grains are measured. This number
of grains is measured because Poldervaart and von Backström (1950)
have shown measurement of any number of grains in excess of 200
crystals does not significantly alter the results. In addition, in mea-
suring a zircon population of this magnitude Greenwood and Greenwood
(1970) have found operator error to be insignificant. As length and
breadth measurements are taken, a visual inspection is made of each
grain to determine if it shows rounded terminations and distortion of
prism faces by overgrowth. Also grains with rounded nuclei, meta-
mict grains, grains with inclusions, and grains with phantom growth
lines are noted. It is found that these features are best visualized by
graphical representation when comparing populations. From visual
inspection it becomes apparent that each zircon concentrate in this
study has its own individual characteristics.

## STATISTICAL METHODS USED IN TREATMENT OF DATA

### Statistical Packages and the Analysis of Zircon Morphology

Most large computer centers have access to collections of stat-
istical programs that can be run by persons without programming know-
ledge and minimal computer experience. A flexible collection of statis-
tical routines, the BMD Biomedical Computer Programs edited by
W. J. Dixon (1971) and developed by the University of California, was
used in this study.
These statistical packages are ideally suited to many types of
geologic data reduction and analysis. In this case, the analysis of zir-
con shapes and sizes is especially amenable to them. Two BMD pro-
grams were used in this study for that purpose. The first (BMD05D)
produces histograms of zircon length, breadth, elongation ratio, size,
and average diameter. This was done by feeding in the raw length and
breadth values for each zircon. By request, the program automatically
does the proper transgeneration on these two variables to calculate
elongation ratio, size and average daimeter of individual zircon crys-
tals. These are calculated as follows:

$$\text{Elongation} = \frac{x_i}{y_i}$$

$$\text{Size} = \sqrt{x_i\, y_i}$$

$$\text{Average Diameter} = \frac{x_i + y_i}{2}$$

where
   $x_i$ = length of zircon crystal i
   $y_i$ = breadth of zircon crystal i

Next the program scales these five measurements and outputs
histograms of the variables. This is the first step used in zircon analy-
sis. Inspection of these histograms gives some idea as to whether each
population is homogeneous (normally distributed) or composed of sev-
eral subpopulations (skewed, overlapping, or multiple distributions).
A better, more objective technique to determine normality is to use the
chi-square goodness-of-fit test procedure described by Davis (1973)
and Griffiths (1967) or the Pearson's beta statistics described by
Griffiths (1967).

A second BMD program (BMD02D) is used to produce scatter
diagrams of zircon length versus breadth and average diameter versus
elongation ratio. Additionally, this program gives sums, means,
standard deviations, and correlation among all the variables. Figures
3 through 8 show a representative output from these programs. Histo-
grams and scatter diagrams of length, breadth, elongation, and aver-
age diameter are presented for a probably single population distribu-
tion, R-3381, and a multipopulation distribution, R-3383.

Scatter diagrams of average diameter versus elongation for the
10 populations in this study show marked differences. Some like R-3381
are very uniform, tight populations; others like R-3383 and R-3385 are
very dispersed and show subpopulations of elongated zircons or large
zircons (see dispersion column in Tables 1 and 2 and Figures 5 and 6).

## Contouring of Zircon Populations

Recent work by Eckelmann (personal communication, 1978) has
shown that zircon suites can be shown visually by contouring a scatter
plot of length and breadth data. The present study indicates that this
is an excellent method for examining for multiple populations. To aid
in this procedure, a small program was written to sum up the number
of zircons in equal size areas on the scatter plot. This data is output
in a format suitable as input to the SYMAP computer mapping program
(Dougenik and Sheehan, 1975). SYMAP uses a standard line printer as
its output device to produce contour maps of the relative number of
zircons falling on any length-breadth location in a sample suite. From
such contour plots, it is generally apparent that a suite of zircons is
homogeneous as in sample R-3381 or is heterogeneous (comprised of
several populations) as in sample R-3383. Figures 9 and 10 show re-
presentative plots for samples analyzed in this study.

## Reduced Major Axis Method

If the histograms, statistics and contour plots just described
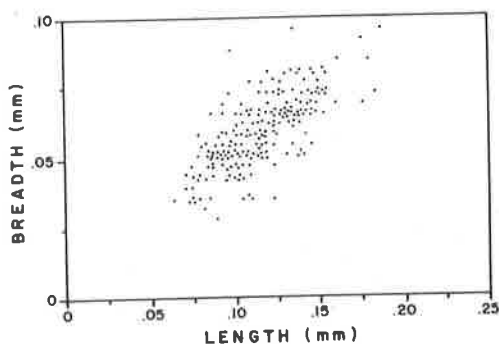
Figure 3. Scatter diagram of length versus breadth for sample R-3381.
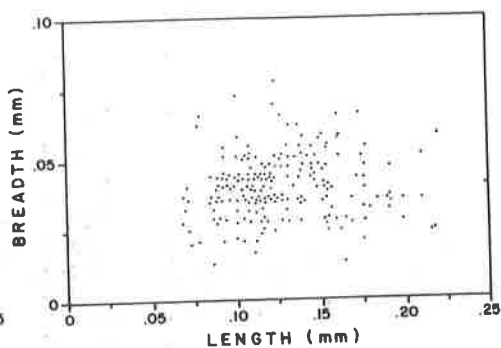


Figure 4. Scatter diagram of length versus breadth for sample R-3383.
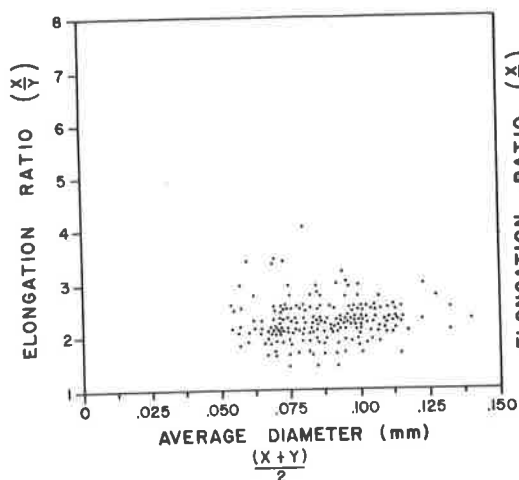


Figure 5. Scatter diagram of average diameter versus elongation of sample R-3381.
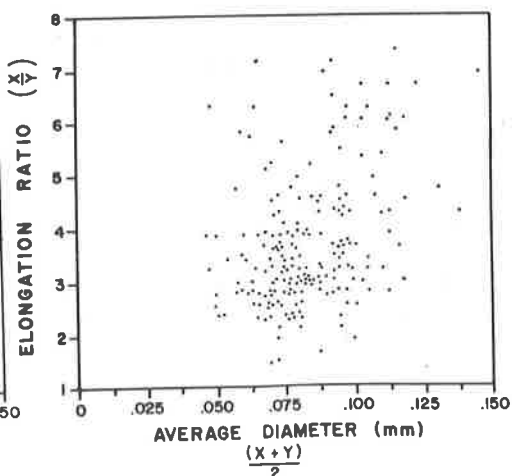


Figure 6. Scatter diagram of average diameter versus elongation for sample R-3383.

indicate that the zircon populations are normally distributed, then the reduced major axis method can be applied to compare and group the zircon populations. If this assumption of normality is false, as it is in this study which indicates mixed populations of metadetrital and igneous zircons (compare Figures 7-8 and 9-10), the statistical calculations are not valid, have unknown meaning and no predictive value as shown by the groupings in Table 3.
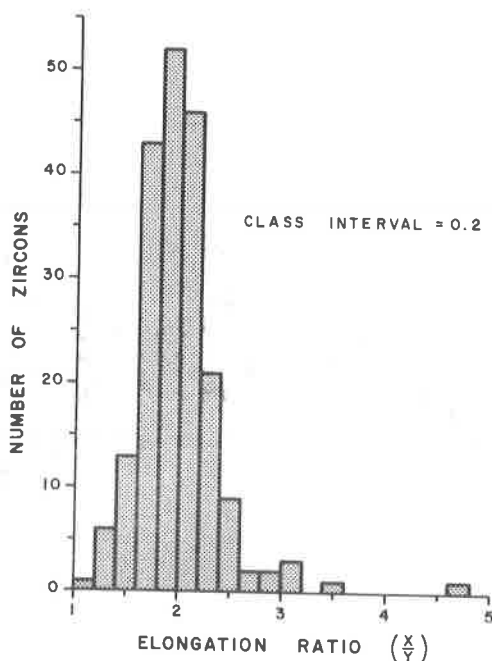
Figure 7. Histogram of elongation ratios for sample R-3381.

Mathematical Basis. The reduced major axis method used to analyze data in this report was developed by Kermack and Haldane (1950) to show growth trends in invertebrate fossils. The method is described in detail by Imbrie (1956). One of the first and most complete discussions of its applications in studying zircons in a suite of igneous rocks is by Larsen and Poldervaart (1957).

In fitting a line to a set of data where error is known to be distributed in both x and y components, the reduced major axis has certain unique advantages. Regression analysis, widely employed as the method of "least squares," has the disadvantage of assigning all error to either the x or the y component; whereas, in reality it is often shared by both. The reduced major axis minimizes the sum of the areas of the triangles formed by lines drawn from each point to the desired line and parallel with x and y axes. Of these two methods, it emerges as the best statistical method for analyzing zircon data because: (1) it does not assume the error resides in only one of the two variables, a significant factor in analyzing scattered data; (2) it does not change with change in scale; and (3) it is simple to compute.

Griffiths (personal communication, 1978) has suggested that regression analysis may be appropriate for zircon analysis and that the reduced major axis need not be used if zircon length and breadth are significantly correlated. More specifically, Griffiths states that the c

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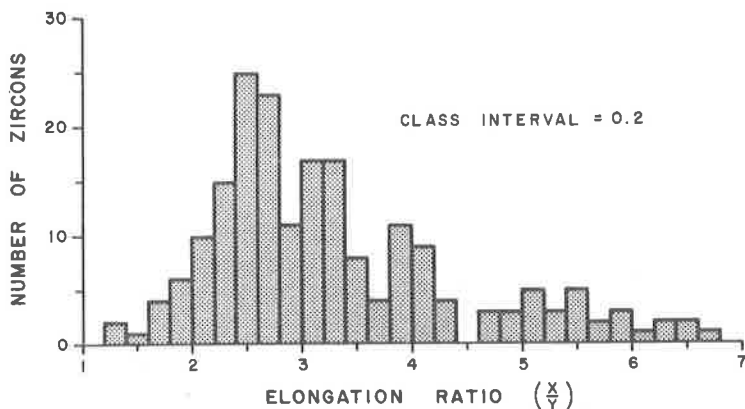
Figure 8.  Histogram  of  elongation ratios for sample
R-3383.
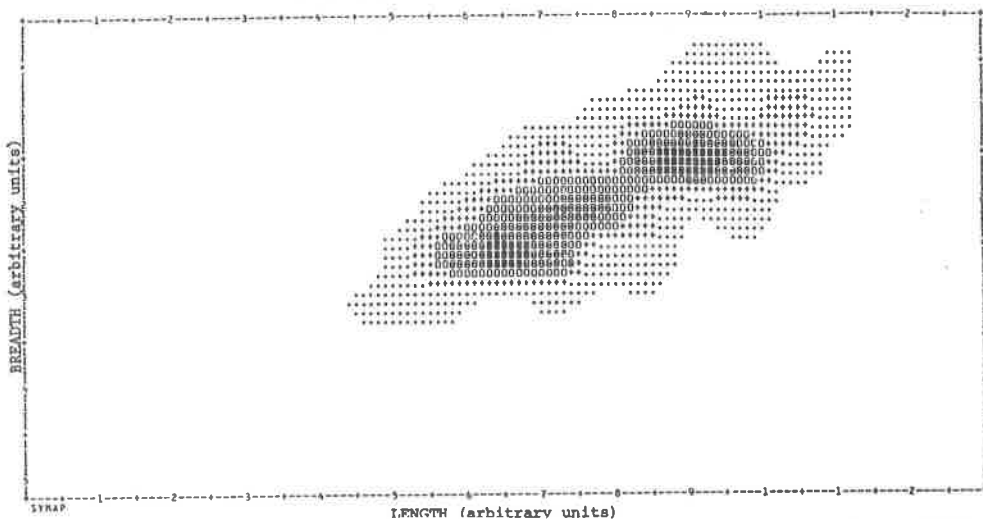


Figure 9.  Contouring of zircon length-breadth data for sample R-3381.

crystallographic axis in  zircon  is  usually  larger than $a_1$ and $a_2$ which
are equal; therefore the lengths of $a_1$ and $a_2$ are, at least partially, de-
pendent on the length of c.   Here form (crystal structure) is determin-
ing.   In each case the length of c may be controlled by the environment
(habit), but lengths of $a_1$ and $a_2$ are still constrained by c (heredity).
In the present study regression analysis is not suitable for all the sam-
ples due to the large scatter of length and breadth data (poor  correla-
tion) among several of the sample populations.
        The basic formulas for the computation  of  the  reduced major
axis are:

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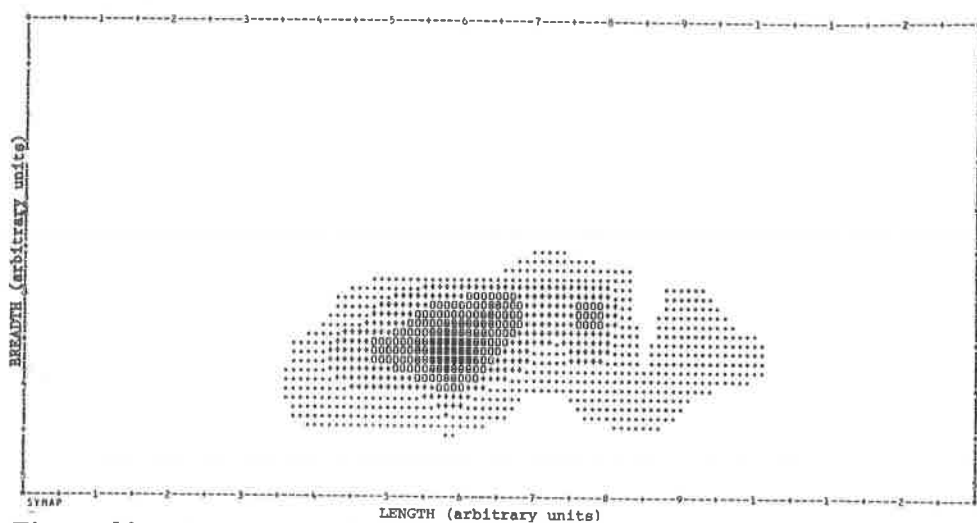
Figure 10. Contouring of zircon length-breadth data for sample R-3383.

$$a = \frac{\sigma_y}{\sigma_x}$$

$$\sigma_a = a\sqrt{\frac{1-r^2}{N}}$$

$$b = \bar{y} - \bar{x}a$$

$$Dd = 100\sqrt{\frac{2(1-r)(\sigma_x^2 + \sigma_y^2)}{\bar{x}^2 + \bar{y}^2}}$$

where

    $a$ = growth ratio or slope

    $b$ = initial growth index or intercept

    $\sigma_a$ = standard error of a

    $Dd$ = coefficient of relative dispersion about the reduced major axis

    $N$ = number of crystals measured

    $\bar{x}$ = mean length of crystals

    $\bar{y}$ = mean breadth of crystals

    $\sigma_x$ = standard deviation of x (length)

    $\sigma_y$ = standard deviation of y (breadth)

    $r$ = Pearson correlation of coefficient

The last six terms listed are all standard statistical measurements found in many texts.

    The problem now arises that, given two or more axes, what is the probability that their differences in slope are real and not due to chance in sampling. This can be calculated from the formula:

$$z = \frac{a_1 - a_2}{\sqrt{\sigma a_1^2 + \sigma a_2^2}}$$

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For large sample populations ($N \geq 200$), if the absolute value of $z$ is greater than 1.96, the probability that the difference in slope arose by chance is less than 0.05. If the absolute value of z is greater than 2.58 the corresponding probability is 0.01. Generally if the probability is less than 0.05 ($z > 1.96$) the hypothesis of equal slopes is rejected, and the observed difference is considered statistically significant. In this case, the position of the lines has little meaning. If however, the probability is greater than 0.05 ($z \leq 1.96$), the hypothesis of identical slopes is usually accepted, and a test is made for the significance of positional differences.

$$z = \frac{x_0 (a_1 - a_2) + (b_1 - b_2)}{\sqrt{\sigma_{a_1}^2 (x_0 - \bar{x}_1)^2 + a_2^2 (x_0 - \bar{x}_2)^2}}$$

As above, if z is greater than 1.96, the difference is taken to be significant at the 5 percent level. If z is less than or equal to 1.96, the observed difference will generally not be accepted as significant. In our calculations we set $x_0 = 10$ for raw data and $x_0 = 1$ for log data.

Several measurements related to the morphology of zircon crystals are included in the output from the computer program presented in this paper. They are elongation, size, and average diameter and are calculated using the same formulas as used by the statistical packages. However, these measurements are averaged and listed for the entire sample population rather than for individual zircon crystals.

Reduced Major Axis Computer Program - A FORTRAN computer program written for this study* calculates reduced major axes through sets of zircon length-breadth data. These axes are then compared with one another using the null hypothesis to determine if statistical difference or statistical identity exist among the sets of data at the 95 percent confidence level.

The program was developed by the second author of this paper to aid in the reduction of large quantities of petrographic data to significant statistical parameters which are more directly comparable. Specifically, this program was used to reduce zircon length and breadth data to the two parameters, slope and intercept by fitting a reduced major axis through the scattered points. In turn, these parameters were statistically compared for different zircon populations to determine into which geologic grouping each zircon population falls. If the premises set forth by Larsen and Poldervaart (1957), Alper and Poldervaart (1957) and Taubernack (1957) are correct then it should be possible to differentiate populations of dissimilar geologic origin from

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*available on request from the authors.

populations of similar geologic origin. This program is not limited to petrographic problems, but is applicable to any study where bivariate analysis can help characterize differences and similarities between samples.

The reduced major axis program is written in FORTRAN IV for a Control Data Corporation Cyber 172 computer. It has a core storage requirement of 15,700 words of 60 bits each. Up to 300 points can be used to determine any axis and up to 50 axes can be computed and compared on any run. Compilation time for this program is approximately 3 seconds and execution time for 20 axes of 200 points each is also approximately 3 seconds. The output of the program is as follows:

1. If requested, the input zircon length and breadth data is listed in tabular form.

2. For each zircon population the following parameters are recorded: mean length and breadth, standard deviation of length and breadth, correlation between length and breadth, slope and intercept of the axis, standard error of the slope, dispersion, elongation, size, and average diameter.

3. All of the computed axes are compared by the z test for significant differences in slope and position. The resultant z values are printed in two tables and zircon populations which show no significant differences are grouped together and listed. Statistical results, using the data in its normal and log-transformed form, are presented in Tables 1 and 2. Kermack and Haldane (1950) and Imbrie (1956) treated their data logarithmically with beneficial results, although Larsen and Poldervaart (1957) did not consider this treatment essential. The computer program here presented is capable of treating the data in either its arithmetic or logarithmically transformed form. The length and breadth data in this study is often positively skewed. This leads to (heteroscedasticity) increasing variance as the magnitude of the axes increase. Because of this, the logarithmic method is judged better than the arithmetic method in studying the relationship between zircon length and breadth as it tends to normalize the skewed data thus reducing heteroscedasticity. It more clearly shows the relationships among the sample populations in bivariate analysis.

Z-tests for comparing slope and position were calculated for all populations at a confidence level equal to or greater than 95 percent. The log transformed data is favored because multiple populations of zircons exist within some of the samples causing them to be heteroscedastic. These subpopulations exert a relatively larger influence on the raw data than they do on the logarithmically transformed data. Thus, the transformed data more realistically portrays the essential character of the main population of zircons from the various rock units.

The groupings resulting from Z-tests for slope and position of the raw and log-transformed data are presented in Table 3. The Z-test, using log data for significant differences in slope and position, indicates that the ten zircon populations can be placed into four groups

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Table 1. Statistical analysis of raw data for zircon populations collected from the Martinsville West quadrangle, Virginia.

| SAMPLE | N | MEAN(X)* | MEAN(Y)* | STD DEV (X)* | STD DEV (Y)* | CORRELATION (R) | SLOPE (A) | INTERCEPT (B)* |
|---|---|---|---|---|---|---|---|---|
| R-3380 | 200 | 50.680 | 18.495 | 11.363 | 4.099 | .337 | .361 | .212 |
| R-3381 | 200 | 46.450 | 23.960 | 9.781 | 5.176 | .711 | .529 | -.630 |
| R-3382 | 200 | 52.675 | 20.645 | 11.469 | 4.451 | .501 | .387 | .239 |
| R-3383 | 200 | 50.975 | 16.330 | 13.245 | 4.401 | .084 | .332 | -.607 |
| R-3384 | 200 | 56.755 | 22.765 | 16.538 | 7.703 | .676 | .466 | -3.670 |
| R-3385 | 200 | 31.965 | 13.405 | 14.781 | 5.703 | .779 | .386 | 1.072 |
| R-3386 | 200 | 47.810 | 19.905 | 11.224 | 4.700 | .644 | .419 | -.113 |
| R-3387 | 199 | 47.809 | 18.809 | 11.914 | 4.480 | .688 | .376 | .829 |
| R-3388 | 200 | 46.730 | 17.360 | 13.113 | 5.962 | .607 | .455 | -3.866 |
| R-3389 | 200 | 44.715 | 20.565 | 10.437 | 4.513 | .648 | .432 | -.931 |

| SAMPLE | STD ERROR -SLOPE | DISPERSION (DD) | ELONGATION | SIZE* | AVE DIAMETER* |
|---|---|---|---|---|---|
| R-3380 | .0241 | 25.761 | 2.844 | 30.357 | 34.588 |
| R-3381 | .0265 | 16.101 | 1.970 | 33.250 | 35.205 |
| R-3382 | .0238 | 21.766 | 2.619 | 32.784 | 36.660 |
| R-3383 | .0235 | 35.296 | 3.360 | 28.405 | 33.653 |
| R-3384 | .0244 | 24.004 | 2.600 | 35.672 | 39.760 |
| R-3385 | .0172 | 30.367 | 2.463 | 20.482 | 22.685 |
| R-3386 | .0226 | 19.831 | 2.451 | 30.702 | 33.858 |
| R-3387 | .0194 | 19.557 | 2.583 | 29.859 | 33.309 |
| R-3388 | .0257 | 25.609 | 2.060 | 28.223 | 32.055 |
| R-3389 | .0234 | 17.742 | 2.460 | 31.648 | 35.140 |

*Values in mm x 400

Table 2. Statistical analysis of log transformed data for zircon populations collected from the Martinsville West quadrangle, Virginia.

LOG TRANSFORMED DATA

| SAMPLE | N | GEOMETRIC MEAN(X)* | GEOMETRIC MEAN(Y)* | GEOMETRIC STD DEV (X) | GEOMETRIC STD DEV (Y) | CORRELATION (R) | SLOPE (A) | INTERCEPT (B) |
|---|---|---|---|---|---|---|---|---|
| R-3380 | 200 | 49.350 | 18.030 | 1.267 | 1.259 | .322 | .973 | -.391 |
| R-3381 | 200 | 45.425 | 23.387 | 1.238 | 1.251 | .704 | 1.050 | -.372 |
| R-3382 | 200 | 51.480 | 20.156 | 1.239 | 1.249 | .485 | 1.040 | -.475 |
| R-3383 | 200 | 49.355 | 15.719 | 1.289 | 1.328 | .118 | 1.118 | -.697 |
| R-3384 | 200 | 54.446 | 21.612 | 1.337 | 1.377 | .669 | 1.101 | -.576 |
| R-3385 | 200 | 29.022 | 12.349 | 1.543 | 1.476 | .734 | .898 | -.220 |
| R-3386 | 200 | 46.552 | 19.391 | 1.260 | 1.255 | .631 | .986 | -.356 |
| R-3387 | 199 | 46.435 | 18.295 | 1.272 | 1.267 | .700 | .984 | -.378 |
| R-3388 | 200 | 44.975 | 16.409 | 1.320 | 1.407 | .587 | 1.231 | -.820 |
| R-3389 | 200 | 48.614 | 20.078 | 1.239 | 1.246 | .661 | 1.027 | -.430 |

| SAMPLE | STD ERROR -SLOPE | DISPERSION (DD) | ELONGATION | SIZE* | AVE DIAMETER* |
|---|---|---|---|---|---|
| R-3380 | .0654 | 7.914 | 2.737 | 29.829 | 33.929 |
| R-3381 | .0530 | 4.802 | 1.942 | 32.594 | 34.514 |
| R-3382 | .0646 | 6.330 | 2.554 | 32.212 | 35.994 |
| R-3383 | .0784 | 10.597 | 3.140 | 27.853 | 32.906 |
| R-3384 | .0561 | 6.468 | 2.519 | 34.303 | 38.267 |
| R-3385 | .0433 | 10.123 | 2.343 | 18.962 | 20.897 |
| R-3386 | .0543 | 5.737 | 2.401 | 30.045 | 33.100 |
| R-3387 | .0501 | 5.422 | 2.538 | 29.147 | 32.475 |
| R-3388 | .0708 | 8.463 | 2.741 | 27.167 | 30.934 |
| R-3389 | .0546 | 5.156 | 2.421 | 31.242 | 34.459 |

*Values in mm x 400

in which all individuals of a group correlate with each other within the prescribed limits of probability. Of the five zircon concentrates from the granite, all correlate with each other within one grouping for both slope and position, except for sample R-3383, which is anomalous. The zircon populations from the gabbro (R-3382), biotite paragneiss

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Table 3.   Significant Groups by Slope and Position.
Log Data

| | | | |
|---|---|---|---|
| R-3383 Δ | R-3380 Δ | R-3380 Δ | R-3380 Δ |
| R-3384 * | R-3385 † | R-3381 † | R-3382 o |
| R-3388 * | R-3386 Δ | R-3386 Δ | R-3384 * |
| | R-3387 Δ | R-3387 Δ | R-3386 Δ |
| | R-3389 Δ | R-3389 Δ | R-3387 Δ |
| | | | R-3389 Δ |

Raw Data

| | | | | |
|---|---|---|---|---|
| R-3381 † | R-3384 * | R-3380 Δ | R-3380 Δ | R-3382 o |
| | R-3388 * | R-3382 o | R-3382 o | R-3385 † |
| | | R-3383 Δ | R-3385 † | R-3386 Δ |
| | | R-3387 Δ | R-3386 Δ | R-3387 Δ |
| | | | R-3387 Δ | R-3389 Δ |

Δ granite
o gabbro
† biotite paragneiss
* pegmatite

(R-3381), and  migmatitic biotite paragneiss (R-3385) all correlate by both slope and  position  with four of the five zircon concentrates from the granite (R-3380, R-3386, R-3387, and R-3389), see Table 3.   This poor grouping  is  probably  due to mixed populations of zicrons within each sample.

Unlike plutons studied by Larsen and Poldervaart (1957), Alper and Poldervaart (1957), and  Taubernack (1957), variations do occur in the zircon populations collected in the granite.  Although any individual will correlate with more than one individual from the rest of the granite, not all individuals will  correlate  with each other, and some correlate with units outside the igneous complex.

The two populations  from  the pegmatite (R-3384) and alaskitic pegmatite (R-3388)  correlate with each other and with the one anomalous sample (R-3383) from the granite.  Additionally, the  pegmatite (R-3384)  shows  the  most  ambiguous relationships of all the samples collected  when  compared  in  all  of its possible groupings.  Not only does it correlate with the  alaskitic  pegmatite but also with all of the granite samples and the gabbro.

Populations of detrital zircons in sedimentary rocks (Poldervaart, 1955, p.  441) and  metadetrital  zircons in metamorphic rocks (Eckelmann and  Kulp, 1956; Gastil and others, 1967) have elongation ratios that do not generally exceed 2. 0.  All zircon populations studied, with the  exception of sample  number  R-3381 (biotite paragneiss), exceed the 2. 0 threshold  established for detrital grains (Tables 1 and 2). The next lowest  elongation  ratio is  from a  zircon population extrated from sample R-3385, migmatitic biotite paragneiss.  The other populations range  from  2. 4 to 3. 4.  Only  one  population (R-3383) from the granite has  an elongation ratio that exceeds 3. 0.  The rest range between 2. 4 and 2. 8.  This  suggests  that all but one population (R-3381,

biotite paragneiss) contain enough zircons of igneous origin to exceed the 2.0 detrital-metadetrital threshold.

## CONCLUSIONS

Visual inspection for rounded grains with overgrowths has been found to be the easiest method of detecting the presence of detrital grains and mixed populations. Next, scatter plots and contour maps of zircon length-breadth data graphically show the character of the sample populations. The reduced major axis method has been considered the best statistical tool for comparing zircon populations (Larsen and Poldervaart, 1957, p. 547). However, present work indicates that it should not be used alone, as serious misinterpretations of the data could result, especially in the study of rocks with complex histories or that may contain mixed populations of both metadetrital and igneous zircons. The reduced major-axis method and other common statistical measures such as mean, standard deviation, and correlation are based on the hypothesis that data follow a normal distribution. If this assumption is false, as it is in the case of mixed populations of metadetrital and igneous zircons (compare Figures 7-8 and 9-10) and other skewed or polymodal distributions, the statistical calculations break down and have unknown meaning and no predictive value. Elongation ratios indicate whether zircons are of igneous or metadetrital origin. The computer programs greatly reduce the amount of time required to mathematically treat the data and thus greatly reduce the labor required to statistically compare populations.

From visual inspection it is apparent that the two samples from the paragneiss as well as the igneous rocks contain metadetrital grains. In addition, the paragneisses also seem to contain zircons that statistically correlate with populations from the granite. The presence of mixed populations of metadetrital and igneous zircons in varying proportions is proposed to be the primary case of the general lack of correlation among populations from the granite.

Elongation ratios suggest that all the populations, with the exception of sample R-3381, biotite paragneiss, are of igneous origin; whereas, visual inspection would suggest that most of these "igneous" populations are composed in part of metadetrital grains. This suggests either that the number of metadetrital grains were not enough to influence the statistical average or the unlikely circumstance that the detrital zircons only acted as nucleation sites for the igneous overgrowths and did not affect length and breadth of the igneous overgrowths.

A study of zircons such as this is extremely helpful in determining the origin of rocks that have had a complex history. It would seem that such an in-depth study would be absolutely essential before trying to interpret radiometric dates derived from zircons, as the

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radioactivity clocks of detrital zircons are rarely reset when they are incorporated in igneous rocks (Gastil and others, 1967). For this reason the determination of the presence of metadetrital grains in an igneous population is necessary for Pb-U age dating as ages determined from such mixed populations would be older than the time of crystallization of the igneous rock (Higgins and others, 1977). Detrital and metadetrital grains as contaminants in igneous populations also cause problems in comparing statistical growth trends among various igneous populations, thus decreasing the usefulness of the reduced major axis method of statistical comparison.

## REFERENCES CITED

Alper, A. M. and Poldervaart, Arie, 1957, Zircons from the Animas stock and associated rocks, New Mexico: Econ. Geology, v. 52, p. 952-971.

Conley, J. F. and Henika, W. S., 1973, Geology of the Snow Creek, Martinsville East, Price, and Spray quadrangles, Virginia: Virginia Div. Mineral Resources Rept. Inv. 33, 71 p.

Conley, J. F. and Toewe, E. C., 1968, Geology of the Martinsville West quadrangle, Virginia: Virginia Div. Mineral Resources Rept. Inv. 16, 44 p.

Davis, J. C., 1973, Statistics and Data Analysis in Geology: John Wiley and Sons, Inc., New York, 550 p.

Dixon, W. J., 1971, BMD Biomedical Computer Programs: Berkley, California, University of California Press, 600 p.

Dougenik, J. A. and Sheehan, D. E., 1975, SYMAP: User's Reference Manual: Laboratory for Computer Graphics and Spatial Analysis, Graduate School of Design, Harvard University, 166 p.

Drummond, K. M., 1962, Zircon studies in the southeastern Piedmont: Unpublished M. S. thesis, Univ. of South Carolina, 26 p.

Eckelmann, F. D. and Kulp, J. L., 1956, The sedimentary origin and stratigraphic equivalence of the so-called Cranberry and Henderson granites in western North Carolina: Am. Jour. Sci., v. 254, p. 288-315.

Gastil, R. G., DeLisle, Mark, and Morgan, J. R., 1967, Some effects of progressive metamorphism on zircons: Geol. Soc. America Bull., v. 78, p. 879-905.

Greenwood, W. R. and Greenwood, J. C., 1970, Experimental error in zircon crystal shape measurements: Jour. Geology, v. 78, p. 576-581.

Griffiths, J. C., 1967, Scientific Method in the Analysis of Sediments: McGraw-Hill, Inc., New York, 508 p.

Hedberg, M. R. and Greenwood, W. R., 1970, Computer analysis of zircon morphology data: Idaho Bureau of Mines and Geology Information Circular No. 21, 15 p.

Higgins, M. W. and others, 1977, U-Pb zircon dates from the central Appalachian Piedmont--A possible case for inherited radiogenic lead: Geol. Soc. America Bull., v. 88, no. 1, p. 125-132.

Imbrie, John, 1956, Biometrical methods in the study of invertebrate fossils: Am. Mus. Nat. History Bull., v. 108, art. 2, p. 215-252.

Kermack, K. A. and Haldane, J. B. S., 1950, Organic correlation and allometry: Biometrika, v. 37, p. 30-41.

Larsen, L. H., and Poldervaart, Arie, 1957, Measurement and distribution of zircons in some granitic rocks of magmatic origin: Mineralog. Mag., v. 31, p. 544-564.

Neuerburg, G. J., 1975, A procedure, using hydrofluoric acid, for quantitative mineral separations from silicate rocks: Jour. Research, U. S. Geol. Survey, v. 3, no. 3, p. 377-378.

Poldervaart, Arie, 1955, Zircons in rocks -- Part 1, Sedimentary rocks: Am. Jour. Sci., v. 253, p. 433-461.

_____, 1956, Zircon in rocks -- Part 2, Igneous rocks: Am. Jour. Sci., v. 254, p. 521-554.

Poldervaart, Arie, and von Backström, J. W., 1950, A study of an area at Kakamas (Cape Province): Geol. Soc. South Africa Trans., v. 52, p. 433-495.

Silver, L. T. and Deutsch, Sara, 1963, Uranium-lead isotopic variations in zircons, a case study: Jour. Geology, v. 71, p. 721-758.

Taubernack, W. H., 1957, Geology of the Elkhorn Mountains, northeastern Oregon, Bald Mountain batholith: Geol. Soc. America Bull., v. 68, p. 181-238.

# BACK CREEK SILTSTONE MEMBER OF DEVONIAN BRALLIER

# FORMATION IN VIRGINIA AND WEST VIRGINIA

By

Katharine Lee Avary*

and

John M. Dennison
Geology Department
University of North Carolina
Chapel Hill, North Carolina 27514

## ABSTRACT

The Back Creek Siltstone, a new Member of the Brallier Formation (Upper Devonian), is found in eleven counties in western Virginia and eastern West Virginia, both in outcrop and in the subsurface. The Back Creek Siltstone has a distinctive stratigraphic position, 50-200 feet (15.3 - 61 meters) above the base of the Brallier Formation and 150-400 feet (45.8 - 122 meters) above the top of the Tully Limestone, and an areal extent of 2,250 square miles (5,830 square kilometers).

Two depositional tongues of Back Creek Siltstone are apparent from the isopachous map. The presence of these two depositional tongues of Back Creek Siltstone, the alternating "d" and "e" Bouma divisions and minor occurrences of "c" division beds, the content-graded beds, the small scale cross-bedding, ripple marks, and scour and tool marks on the bases of siltstone beds all support the idea of a turbidite origin for the Back Creek Siltstone. More specifically, the environment of deposition of the Back Creek Siltstone is interpreted as the outer portion of a submarine fan complex. The Back Creek Siltstone outer fan is stratigraphically the oldest deposit associated with the Augusta Lobe of the Catskill delta complex. The distribution of the Back Creek Siltstone may have been controlled by syndepositional growth structures along the north margin of the area of occurrence.

The Back Creek Siltstone may have enough open-fracture porosity for gas reservoir potential; near the eastern limits of occurrence, where it becomes as coarse as fine sand, it may have suitable intergranular porosity.

*Present address: West Virginia Geological and Economic Survey, P. O. Box 879, Morgantown, West Virginia 26505.
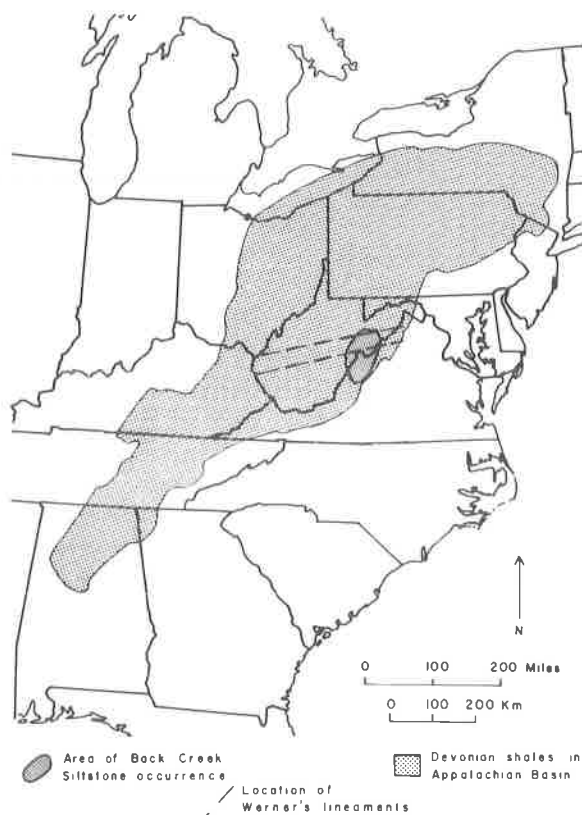
Figure 1. Index map showing the area of occurrence of the Back Creek Siltstone, the extent of the underlying Devonian shales in the Appalachian Basin, and the lineaments recognized by Werner (1979).

## INTRODUCTION

The Back Creek Siltstone Member is herein proposed as a formal name for a distinctive and useful stratigraphic unit in the lower Brallier Formation (Upper Devonian) in western Virginia and eastern West Virginia (Figures 1 and 2). The Back Creek Siltstone was recognized along the Allegheny Front and informally named by Dennison (Dennison and DeWitt, 1972; Dennison and Hasson, 1977).

The Back Creek Siltstone is a sequence of light olive gray, thinly to thickly bedded siltstones, interbedded with light olive gray, thinly laminated to thickly bedded shales and silty shales. The siltstone beds are distinguished in outcrop from typical Brallier Formation siltstones because of the thickness and abundance of siltstone relative to interbedded shale. Often the beds are fractured; some of the fractures are
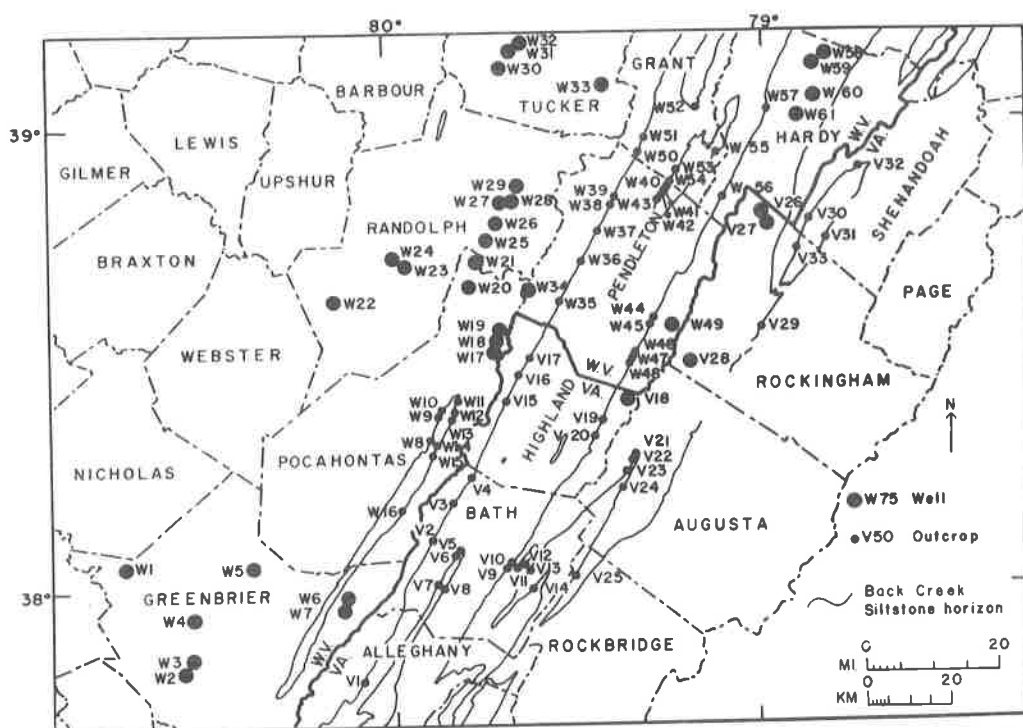
Figure 2. Location map of the Back Creek Siltstone outcrop and well data points. Detailed location descriptions of each data point are in Appendix 1 in Avary (1978).

filled with quartz and calcite.

The known extent of the Back Creek Siltstone is an area of 2,250 square miles (5,830 square kilometers) in eleven counties.

This three-dimensional stratigraphic study of the Back Creek Siltstone shows that it has a relatively constant stratigraphic position above the base of the Brallier Formation and the top of the Tully Limestone, and thus is probably a useful time-marker.

The Back Creek Siltstone serves as the first record of the Augusta Lobe of the Catskill delta and is interpreted to be a turbidite, deposited on the outer fan portion of a submarine fan complex.

The present paper is a combination of the thesis by Avary (1978) and the earlier observations of Dennison, not published in detail.

## Location

The Back Creek Siltstone has been identified at 51 outcrops in Augusta, Bath, Highland, Rockbridge, Rockingham, and Shenandoah Counties, Virginia, and in Grant, Hardy, Pendleton, and Pocahontas Counties, West Virginia. In addition, the Back Creek Siltstone has

been located in 15 wells drilled in Highland and Rockingham Counties, Virginia, and in Pendleton, Pocahontas, and Randolph Counties, West Virginia (Figure 2). The area of occurrence of the Back Creek Siltstone is 75 miles (120 km) along the Appalachian outcrop trend and 30 miles (48 km) across the strike belts.

Almost all Back Creek Siltstone outcrops are road cuts, although a few exposures are in stream beds and roadside quarries. Well data is in the form of commercially prepared lithologic and gamma-ray logs. (Appendix 1 in Avary (1978) contains a complete listing of data points used in the study of the Back Creek Siltstone.)

## Geologic Setting

The Back Creek Siltstone occurs in the lower portion of the Upper Devonian Brallier Formation and is thus Finger Lakes Age. The Back Creek Siltstone was deposited about 358 million years ago, based on its position in the Central Appalachian stratigraphic column (Figure 3). The Back Creek Siltstone can be positioned in terms of physical stratigraphy by establishing its distance above the base of the Brallier Formation, above the top of the Tully Limestone Member of the Millboro or Harrell Shales, above the top of the Purcell member of the Marcellus Shale or above the top of the Tioga Bentonite. All of these horizons can be traced over large areas of the Appalachian basin, and, except for the base of the Brallier Formation, they appear to serve as consistent time-markers. The base of the Brallier Formation is probably a somewhat diachronous facies boundary, becoming younger to the west and southwest across Figures 2 and 3. The Back Creek Siltstone is a small part of the large influx of clastic material which comprises the Catskill delta complex.

The outcrops of the Back Creek Siltstone are in the Valley and Ridge physiographic province and in the most eastern outcrop belt of Upper Devonian rocks (Browns Mountain anticline) in the Allegheny Plateau. The Allegheny Front serves as the boundary between the two provinces and provides a series of excellent exposures of the Back Creek Siltstone in the homoclinal outcrop belt at the base of this escarpment. The wells which penetrate the Back Creek Siltstone in the western part of the area are located in the Allegheny Plateau.

## Methods

At each outcrop, Avary made detailed measurements of the siltstone beds and the intervening shale beds. The bed thicknesses were measured to the nearest centimeter and the data recorded and plotted on columnar sections. At every outcrop where possible, stratigraphic control was obtained using the distance from the base of the Brallier Formation to the base of the Back Creek Siltstone, and/or the distance between the top of the Tully Limestone and the base of the Back Creek
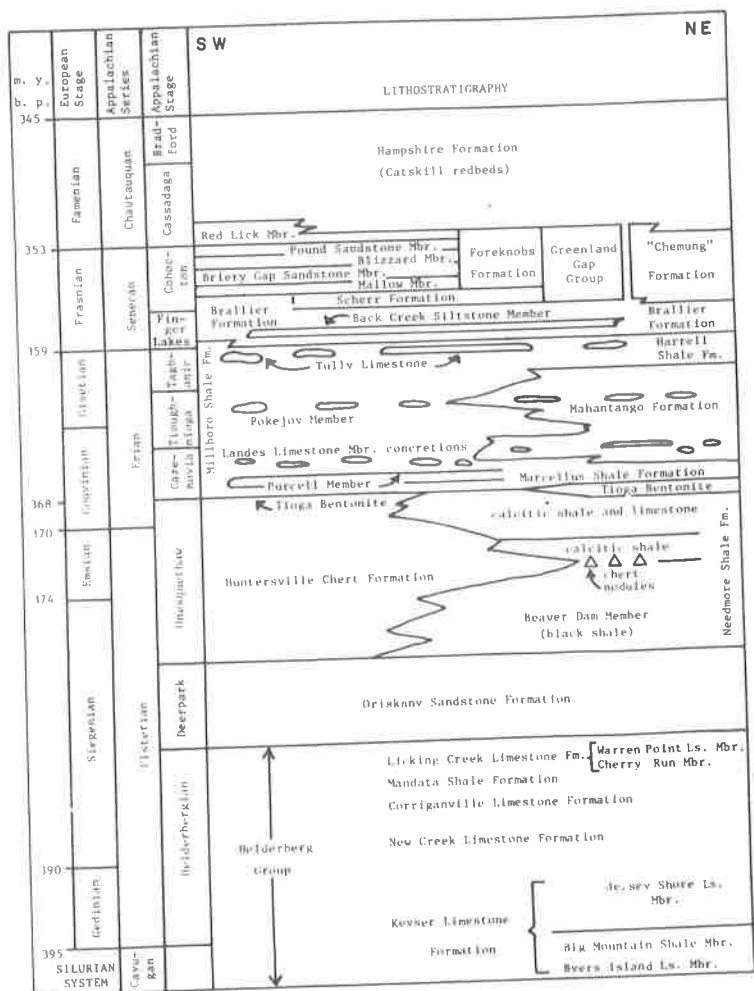
Figure 3. Stratigraphic nomenclature chart of the Devonian in western Virginia and eastern West Virginia. Radiometric age are from Harland, Smith and Wilcox (1964).

Siltstone. Control for these stratigraphic distances at many outcrops was based on previous field work and plane table data by Dennison in conjunction with Kenneth O. Hasson, Orville D. Naegele and Jack W. Travis.

In addition to measurements and descriptions of bed thickness using Ingram's (1954) terminology, the presence of sole markings and the paleocurrent data obtainable from them were recorded. Samples were collected for thin-sectioning, and many of the outcrops were photographed. The field data were used to generate a series of maps showing

the total thickness of the Back Creek Siltstone, the thickness of the thickest siltstone bed, the percentage of siltstone, the distance between the base of the Brallier Formation and the base of the Back Creek Siltstone, and the distance from the top of the Tully Limestone to the base of the Back Creek Siltstone. All of these maps (Figures 4, 9, 10, 11, and 12) were drawn on a palinspastic base map, which was constructed using the palinspastic map of the Onesquethaw Stage (Dennison, 1960, 1961) as a guide for restoring the Back Creek Siltstone outcrop trace.

The well data were analyzed by comparing lithologic and gamma-ray logs. Packets of siltstones and very fine sandstones described on the lithologic logs at the appropriate stratigraphic position were correlated with "kicks" on the gamma ray logs which indicated lower concentrations of radioactivity and hence less shaly beds. The Tully Limestone, the Pokejoy Member limy beds, Clearville siltstone interval, the Purcell member, and the Tioga Bentonite were all used for stratigraphic control and correlation between wells. (The variation in capitalization of stratigraphic names reflects the standard stratigraphic notation of capitalization of formally established stratigraphic names and lack of capitalization for informal names.)

## Previous Work

The Back Creek Siltstone outcrop has not been mapped on any published areal geologic map. Dennison and Naegle (1963, p. 21) made the first description of what is now called the Back Creek Siltstone. "Conspicuous siltstones occur about 60 feet above the base of the Brallier," in the Allegheny Front outcrop belt in Pendleton County. They also suggested that the Brallier Formation is locally tectonically thinned in Pendleton County, as evidenced by offsets in the siltstones.

Dennison (1965, p. 297) first suggested a Devonian clastic source east of Staunton, Virginia beginning during Brallier Formation deposition, and reflected in the prominent lower Brallier siltstones and some reddish intertongues near the top of the Brallier Formation in Highland County, Virginia.

Kulander (1968, p. 37-38) recognized basal massive siltstones within the Brallier Formation in his mapping of the northern half of Browns Mountain anticline in Pocahontas County, West Virginia.

The clastic source with input into the Devonian seaway was named the Augusta Lobe of the Catskill delta by Dennison (1970, p. 66) and the Back Creek Siltstone was shown as an unnamed gray siltstone member in the Brallier Formation (Dennison, 1970, Figure 3). Similar stratigraphic cross-sections were shown by Dennison (1971, Figure 6) and McGhee and Dennison (1976, Figure 2). Also, Dennison (1970, p. 64) noted the erroneous mapping of the massive siltstones in the lower Brallier Formation as basal "Chemung" Formation near Mouth of Seneca (W-38 and W-39) in Pendleton County, West Virginia by Tilton, Prouty, Price and Tucker (1927).

Dennison (1971, p. 1181) named Grant Bay between the Fulton and Augusta Lobes of the Catskill Delta. Also in that paper, the first suggestion is made of a turbidite origin for these siltstones (Dennison, 1971, p. 1187).

The first map showing the location of the Augusta Lobe and Grant Bay at the base of the Cohocton Age was published by Dennison and DeWitt (1972, Figure 34). Dennison and DeWitt (1972, p. 112) state: "A prominent turbidite siltstone a few tens of feet above the base of the Brallier probably represents another time-surface clastic pulse which can be correlated in both outcrop belts (Allegheny Front and Browns Mountain anticline), a unit Dennison informally refers to as the Back Creek siltstone." This was the first use of the name Back Creek siltstone in print. The name was reserved with the Geologic Names Committee of the United States Geological Survey for formal publication of a new stratigraphic unit, and the present publication constitues the formal introduction of the Back Creek Siltstone Member of the Brallier Formation.

Another map (Dennison and Head, 1975, Figure 12) showed the location of the Augusta Lobe relative to the Fulton, Snyder, and Wyoming Lobes (Willard, 1939) at the beginning of the Cohocton Age (base of Scherr and "Chemung" Formations in Valley and Ridge outcrop belts).

In a stratigraphic cross-section of Devonian shales along the Allegheny Front, Dennison and Hasson also show the Back Creek Siltstone (Dennison and Hasson, 1979; Hasson and Dennison, 1978, Figure 4). In addition, they describe the Back Creek Siltstone briefly and again state that the Back Creek Siltstone is a probable time-marker and that it "apparently originated as turbidity slumps off the Augusta Lobe of the delta complex in Virginia."

# STRATIGRAPHY

## General Stratigraphy of the Brallier Formation

The Brallier Formation was named by Butts (1918) for exposures at Brallier Station on the Huntington and Broad Top Mountain Railroad, five miles northeast of Everett, Bedford County, Pennsylvania.

Butts (1933) adopted the name Brallier for use in Virginia, and in 1932 he had personally communicated his recommendation to Woodward who was then doing geologic work in the Roanoke, Virginia area (Woodward, 1932, p. 63).

The name Brallier Shale was adopted by Woodward (1943) for use in West Virginia, replacing earlier names of Portage Shale, Woodmont Shale, and Jennings Formation which had variously been applied to strata now known as Brallier Formation.

The Brallier Formation consists of "medium-dark gray, thickly laminated shales with distinctly-bounded siltstones in beds as much as 1 foot (0.3 meters) thick and showing evidence of turbidity current origin. In outcrops and wells, the basal contact of the Brallier is placed at the lowest siltstone overlying darker shales of the Harrell or Millboro Shales" (Dennison and Hasson, 1976, p. 279). Lundegard, Samuels, and Pryor (1978) have documented evidence that the siltstones in the Brallier Formation as a whole over the Appalachian basin are of turbidite origin, with the interbedded mudstones, claystones, and shales having a mixed hemipelagic origin.

## Stratigraphy of the Back Creek Siltstone

The Back Creek Siltstone occurs as a Member of the Brallier Formation in eleven counties in western Virginia and eastern West Virginia. The proposed type section for the Back Creek Siltstone is in a road cut along the northeast side of U. S. Route 250, near the valley of Back Creek in the western part of Highland County, Virginia (locality V-16 on Figure 2). A description and map of the proposed type section are included in the Appendix.

An isopachous map (palinspastic base) of the total thickness in feet of the Back Creek Siltstone Member of the Brallier Formation is shown in Figure 4. The zero isopach line is drawn between the points where Back Creek Siltstone is present and points where there is no Back Creek Siltstone. In the field, an arbitrary thickness of some siltstone beds at least 10 cm thick was used to establish the presence of the Back Creek Siltstone. Beyond the limits of this arbitrary cut off of 10 cm thick beds for recognizable Back Creek Siltstone, these particular siltstone influxes could not be distinguished specifically from the other thin siltstone beds in the Brallier Formation. The Back Creek Siltstone is not all siltstone, but contains a large percentage of interbeds of medium dark gray, thickly laminated shale, which weathers light olive
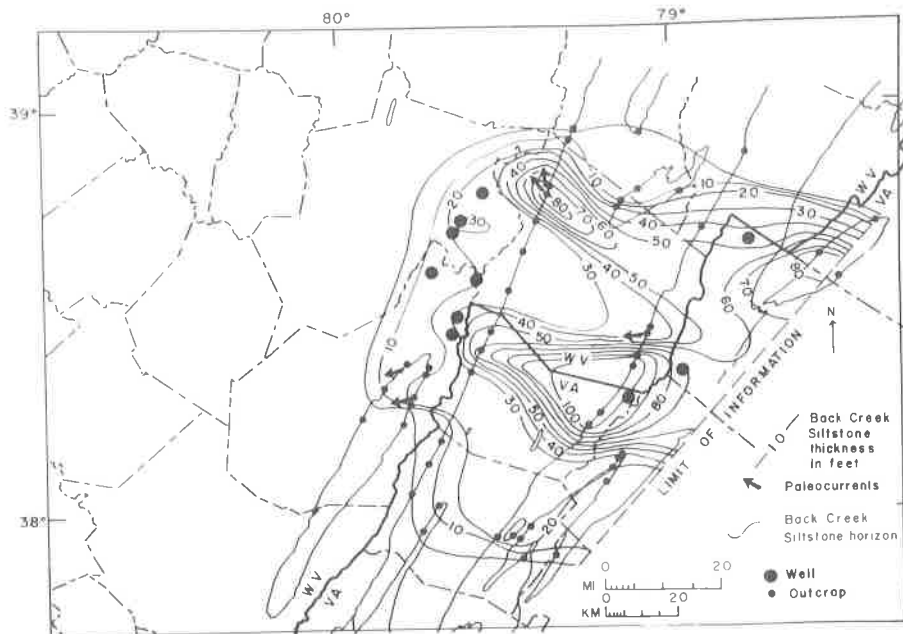
Figure 4. Isopachous map of the total thickness in feet of the Back
Creek Siltstone. (Palinspastic base during Devonian
deposition. Rectangle shows present-day dimensions of
study area.)

gray and chippy. The percentage of shale increases outward from the
eastern source area in the Back Creek Member (Figure 11). In all of
the Back Creek outcrops, prominent siltstones characterize the Mem-
ber, even if locally they constitute less than 50 percent of the total
thickness.

Two depositional tongues separated by an area of thinner sedi-
ment are well shown on the isopachous map. These two tongue-shaped
areas probably represent outer fan depositional lobes extending out
from the general site of the Augusta Lobe of the Catskill delta.

The thickness of the Back Creek Siltstone reaches a maximum
(112 feet or 34 meters) in the east central portion of the area, at Sugar
Grove, Pendleton County (W-46), probably closest to the source of
these sediments. An eastern source area is also indicated by the limit-
ed paleocurrent data obtained from flute and groove casts, as well as the
general Upper Devonian sedimentation patterns.

Figure 5 locates three stratigraphic cross-sections of the Back
Creek Siltstone, which also show the stratigraphic framework of the
underlying Middle Devonian dark shales of the Hamilton Group. These
cross-sections (Figures 6, 7, 8) are all drawn with present day dis-
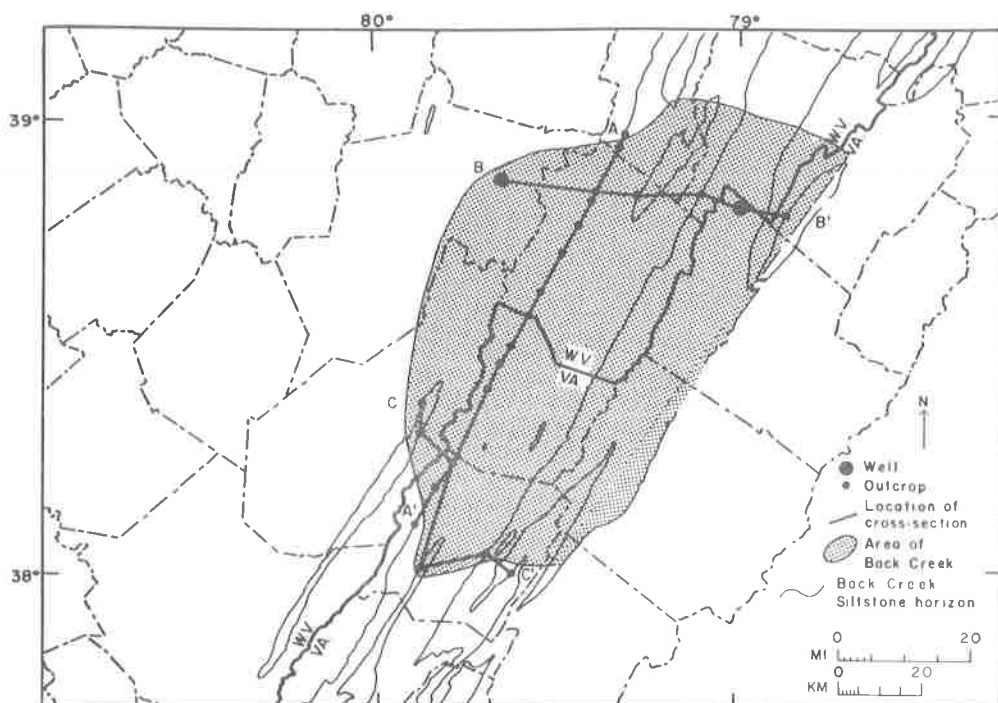tances between data points, not palinspastically restored distances.

Figure 5.  Locations  of three stratigraphic cross-sections showing the position of the Back Creek Siltstone.


In the northeast-southwest section  along  the  Allegheny  Front (A-A', Figure 6),  the  Tully  Limestone concretion zone is used as the horizontal datum;  the top of the Tully Limestone is generally regarded as the boundary between  the  Middle and  Upper  Devonian  (Oliver and others,  1967, 1969;  Rickard,  1974),  and  serves  as an approximate time-marker from the  northern portions of Virginia and West Virginia into southern Pennsylvania.  In southern Pennsylvania, the Tully Lime- stone  becomes thick enough to map as a continuous,  bedded  limestone which  can be  traced  through the subsurface northward into New York (Heckel, 1969).  In the  northeastern  portion of the cross-section, the Brallier Formation is underlain by the Harrell Shale, while in the south- west, the Brallier Formation is underlain by the black to grayish-black Millboro Shale.  The nomenclature change is caused by a facies change; to the northeast, the  dark  gray,  coarser  detrital clastic Mahantango Formation (shown  in  Figure 6 with its Clearville member and Pokejoy Member named)  divides  the  dark  shales into the upper Harrell Shale and the lower Marcellus Shale.  Where the Mahantango Formation is no longer distinguishable,  the  name  Millboro Shale is used for all of the rocks above the Tioga Bentonite and below the base of the Brallier For- mation.  The Purcell  member  and  the  Tioga Bentonite are shown as stratigraphic  reference  horizons.  These approximate time-markers
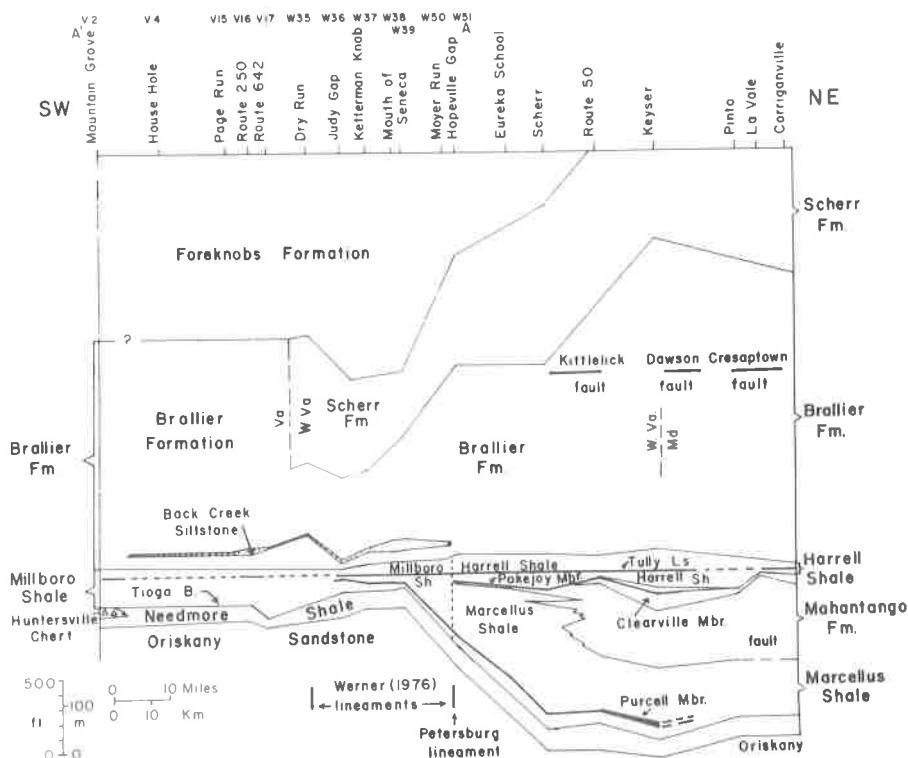
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Figure 6.   Stratigraphic cross-section along the Allegheny Front,
(A-A').

are useful for recognizing the stratigraphic convergence toward the southwest which affects these Devonian rocks along the Allegheny Front.

This cross-section (Figure 6) transects the two depositional tongues of thicker Back Creek Siltstone separated by an area of thinner sediment. Also, the abrupt nature of the north and south margins of the Back Creek Siltstone is apparent.

The northern of the two east-west cross-sections (B-B', Figure 7) cuts across the northern depositional tongue of the Back Creek Siltstone. The Tully Limestone is used as the horizontal datum. The same nomenclature changes for the dark shales below the Brallier Formation apply here as in the Allegheny Front cross-section (A-A'). The Harrell Shale, Mahantango Formation, and Marcellus Shale change facies westward into the Millboro Shale. The Purcell member and the Tioga Bentonite horizons are shown for reference. Tectonic thickening of the strata is apparent in the Randolph Permit 89 well (W-29), as is indicated by the fact that all of the formations are thickened by roughly the same proportion.

The southern of the two east-west cross-sections (C-C', Figure

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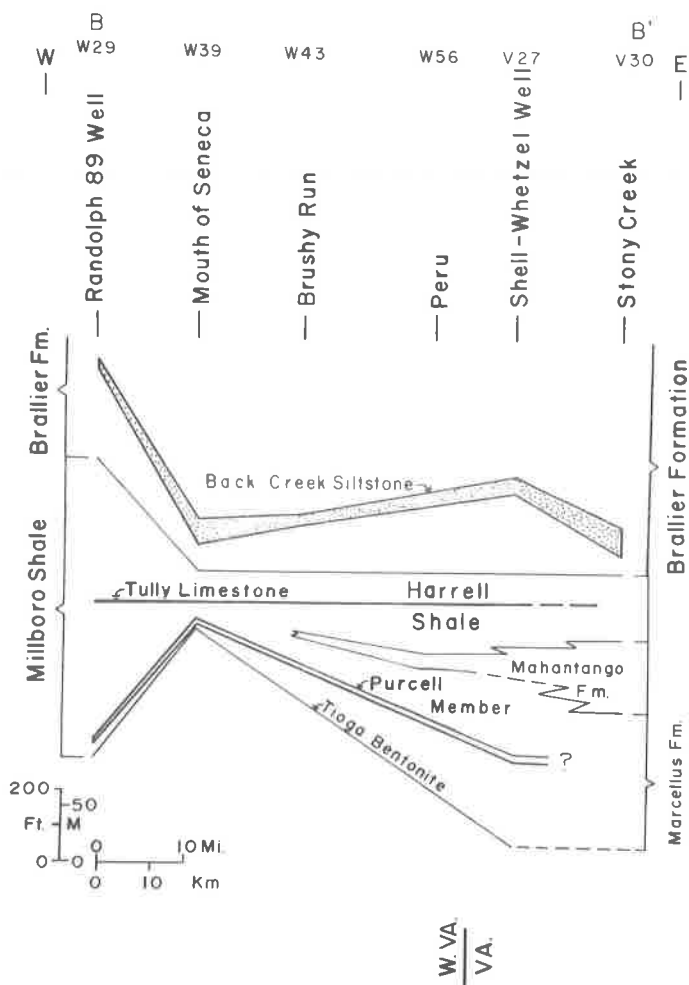
Figure 7. Northern east-west stratigraphic cross-sections, (B-B').

8) shows much the same relationships as the other two cross-sections. The section is hung on the Tully Limestone horizontal datum. The area traversed by this cross-section is south of the Mahantango Formation coarser clastic influence, so all of the dark shale beneath the Brallier is designated Millboro Shale. The same reference horizons, the Purcell member and the Tioga Bentonite, are shown. This section is drawn along the south margin of the southern depositional tongue of the Back Creek Siltstone and can be compared with the thicker Back Creek Siltstone on the northern cross-section (Figure 7).

To further clarify the stratigraphic position of the Back Creek Siltstone, two maps (Figures 9 and 10) were constructed, showing only

W    C
|    W9

Brallier Fm. — Curry Road

Millboro Shale

M
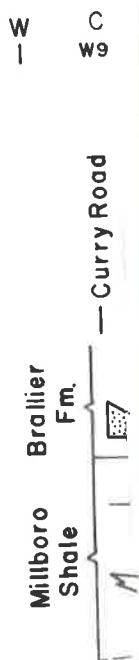
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Figure 8.

those data points which c
    Figure 9 is a map
tance between the base of
Back Creek Siltstone.  Th
bably due to the general cl
of the Brallier Formation i
the detrital clastic source
eastern Randolph and north
cause of the marked thickne
closest outcrop.
    Figure 10 is a  map  (palinspastic  base)  of  the  stratigraphic

Figure 9. Map of the distance in feet between the base of the Bral-
lier Formation and the base of the Back Creek Siltstone.
(Palinspastic base.)

distance between the top of the Tully Limestone concretion zone and the
base of the Back Creek Siltstone. Increased thickness of the interval
in areas corresponding to the two depositional tongues is well-express-
ed by these isopachous lines. The general eastward thickening of the
Tully Limestone to Back Creek Siltstone interval is in agreement with
the Upper Devonian facies patterns which suggest clastic wedge diver-
gence toward an eastern source area.


PETROLOGY

Siltstone Content

The percentage of siltstone was computed by totaling the centi-
meters of siltstone thickness in each measured section and calculating
what percentage of the Back Creek Member is siltstone. These data
are plotted on the map in Figure 11 (palinspastic base). The suggestion
of the two depositional tongues is also seen on this map; however, the
axes of these tongues do not correspond to those of the tongues on the
isopachous or siltstone bed thickness maps (Figures 4 and 12). The
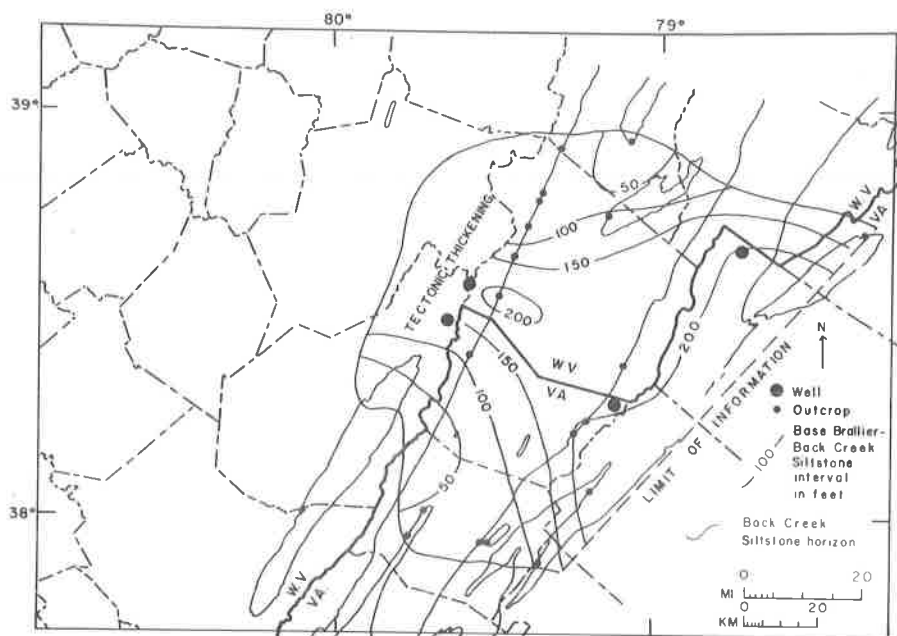concentration of silt to the south of the maximum thickness of the
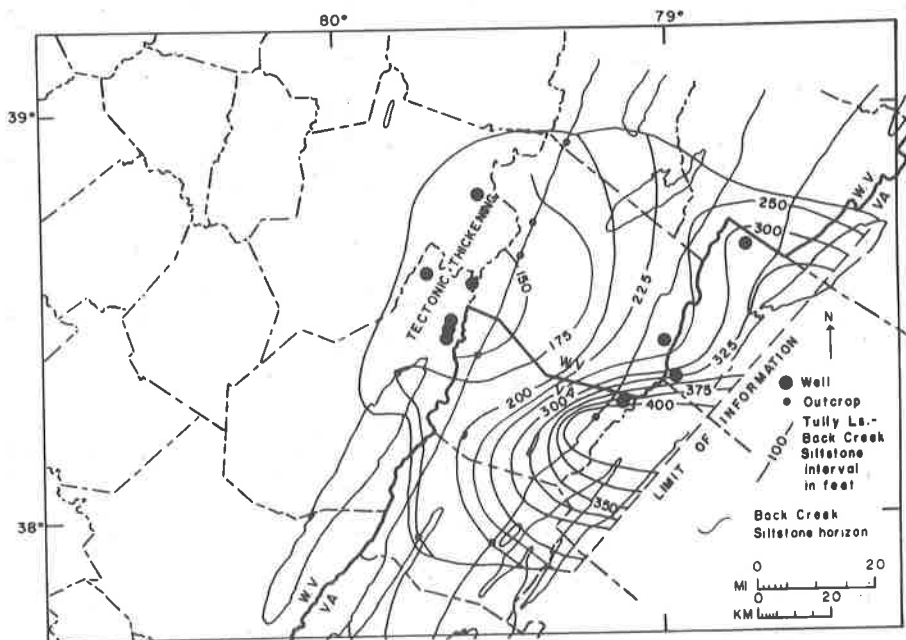
Figure 10. Map of the distance in feet between the top of the Tully Limestone and the base of the Back Creek Siltstone. (Palinspastic base.)

Member, as well as to the south of the maximum bed thickness, may result from a concentration of silt marginal to the main directions of the turbidity flows which formed the tongues seen on the thickness maps. The overall higher percentage of silt in the southern part of the area suggests that perhaps the source area to the south was being eroded more rapidly than the source areas supplying sediment to the northern tongue. Another possible explanation would be that more silt was trapped on the upper portions of the fan and/or up on the delta in the north. The idea that much of the coarser material can remain trapped up on a delta was suggested by McIver (1970, p. 80).

### Thin Section Petrology

Fourteen thin sections of siltstones cut perpendicular to bedding were examined. The Back Creek Siltstone can be described as a mica-ceous coarse siltstone (according to Picard's (1971) classification). The Back Creek siltstone falls into Folk's (1974) sublitharenite category with 80-90% of the grains identified as quartz, 5-10% as rock fragments, and 5-10% as plagioclase.

In addition to these "essential" (according to Folk (1974)) grains which constitute about 40% of the total rock, the Back Creek Siltstone

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Figure 11.   Map of the percentage  of siltstone in the Back Creek
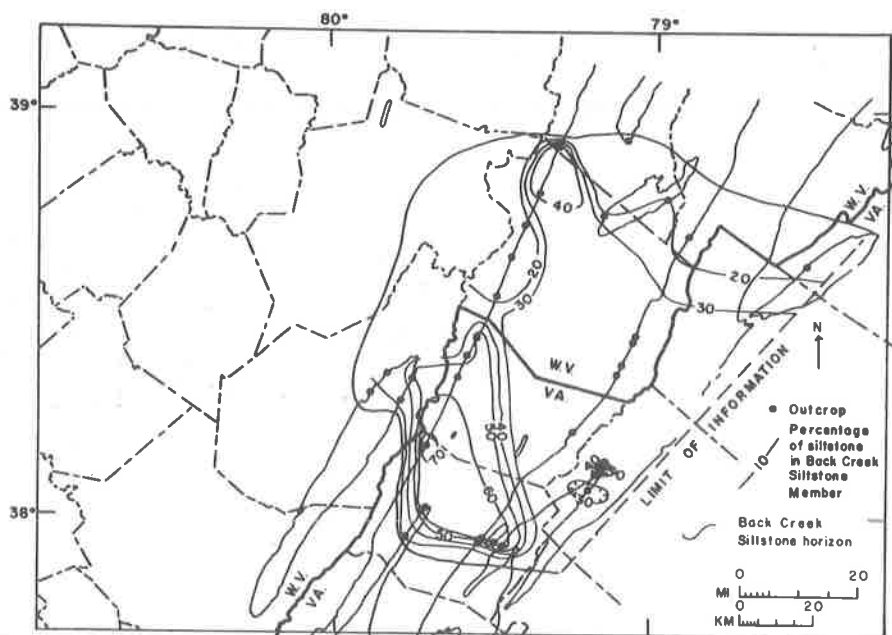              Siltstone Member.   (Palinspastic base. )

contains about 10% mica  and  about  50% matrix.   The mica is largely
muscovite, with some chlorite.   The matrix consists of clay,  probably
some crushed mica grains, very small quartz grains, rock fragments,
plagioclase grains, and some limonite.

        The quartz grains  are generally subangular to subrounded, and
range from  equant to tabular in shape.   The siltstones are moderately
well-sorted, with  about 80 percent of the grains measured falling into
the medium and coarse silt size ranges.   The  Back Creek Siltstone is
classified as an  immature  sediment (Folk, 1974)  on  the basis of its
clay and mica content.

        The present-day  permeability  of the Back Creek Siltstone  is
largely fracture permeability.  Some of the fractures are open, while
others have been filled either partially or completely with quartz.

        Other important features  noted  in the thin sections of the Back
Creek Siltstone are the concentrations of mica grains in roughly paral-
lel alignment near the tops of several beds, in contrast with only scatter-
ed mica grains near the bases of these same beds, and a faint sugges-
tion of  preferred  orientation  parallel  to  bedding  of the long axes of
some quartz silt grains.

        The maximum grain diameter recorded for the Back Creek Silt-
stone is 0. 16 mm.   The most common size range is 0. 02 - 0. 10 mm.

        The study of fine-grained  sedimentary  rocks in thin section is

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difficult due to the tiny, angular nature of the grains and the abundance of matrix. Part of the problem is that many of the particles are smaller than the 0.03 mm thickness of the thin section. Bouma and Hollister (1973, p. 102) state: "Silt and fine sand grains are not visible when scattered through a clay matrix. Consolidation moves these grains closer together and when a certain density is reached they become visible as a thin layer."

Many workers have questioned the origin of the matrix of "graywackes." Several people believe that large amounts of matrix are of secondary origin. Kuenen (1966), on the basis of his experimental work on the matrix of turbidites, concluded that much of the matrix of graywackes is the result of pressure solution processes. The objective of our study is not to discover the origin of matrix in graywackes; however, it is necessary to be aware of the many problems that arise in the study of fine-grained rocks with abundant matrix.


## SEDIMENTOLOGY

### Bedding of Siltstones

The thickness in feet of the thickest Back Creek siltstone bed measured at each outcrop is shown on the map in Figure 12 (palinspastic base). The bed thickness decreases toward the west, away from the maximum thickness of 5.2 feet (1.7 meters) measured at Panther Run, Pendleton County (W-48). The pattern of depositional tongues is apparent on this map and reflects the channeling of most of the available sediment into two distinct areas on the outer fan-shaped region.

Numerous other studies of turbidites have also documented a sourceward increase in bedding thickness of the sandstones or siltstones interbedded with shales (Enos, 1969; Schenk, 1970; Bennetts, 1974; and Kepferle and others, 1977).

### Internal Sedimentary Structures

Graded bedding has been recognized by many workers as one criterion for recognizing sediments deposited by turbidity currents. The Back Creek siltstone beds are graded in the sense that many of them have shaly tops expressed as irregular upper contacts, in contrast to sharp, regular bottom contacts. However, the beds do not exhibit a complete systematically sorted range of grain sizes. The suggestion of upward decrease in grain size is noticeable in thin section where concentrations of mica grains with roughly parallel alignment are found at the top of a bed and not at its base. The lack of dramatic grading of siltstone beds is due to the relatively uniform grain sizes suspended in the turbid waters which deposited the Back Creek Siltstone. This lack of pronounced grading has been noted in other siltstones interpreted to
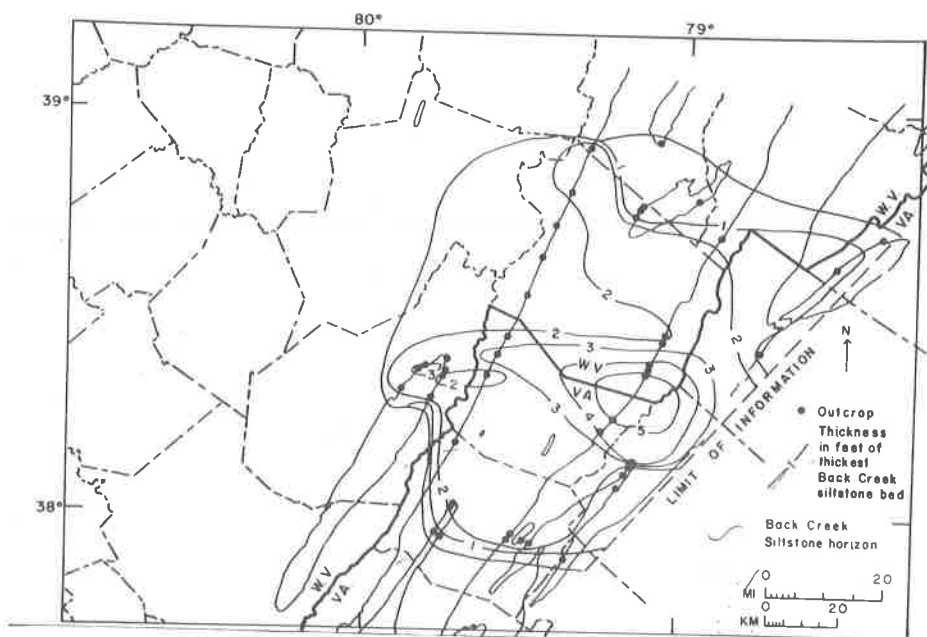
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Figure 12. Map showing the thickness in feet of the thickest silt-
stone bed in the Back Creek Siltstone Member. (Palin-
spastic base.)

be turbidites. McBride (1962, p. 50) refers to beds with increasing
amounts of shale upward as "content graded bedding" in his study of
the Martinsburg Formation (Ordovician) in the central Appalachians.
McIver (1970, p. 76) notes a lack of graded bedding in the "Portage"
rocks he studied (equivalent stratigraphically to the Brallier Formation
of the Virginias). Frakes (1967, p. 65-69) detected slight grading in
thin sections of the Trimmers Rock Member of the Fort Littleton For-
mation (Upper Devonian) in eastern Pennsylvania.

The shaly tops of the beds of the Back Creek Siltstone are pro-
bably due to an upward relative increase in clay content over silt con-
tent. These changes may reflect a decrease in the number of larger
silt particles available from the source area. Probably, decreasing
current energy is a more likely cause, since these shaly tops are fairly
common. The source area would have to have changed periodically and
abruptly to produce the shaly tops of otherwise coarser beds; such
source area changes seem less reasonable.

At some outcrops, the presence of small-scale cross-bedding
of Back Creek siltstone beds was noted. The cross-bedding is a
characteristic feature of the "c" division of Bouma's (1962) turbidite
sequence; the "c" division represents more proximal turbidite deposits.

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## External Sedimentary Structures

Two varieties of bedding surface sedimentary structures are recorded from the base of Back Creek siltstone beds; these are both current structures (Pettijohn and Potter, 1964, p. 7). Scour marks occur in the form of flute casts on some beds. Tool marks, chiefly groove casts, were found on the base of a few beds. These structures are not abundant at the outcrops of the Back Creek Siltstone examined. Bedding plane exposures are uncommon. Probably, the apparent scarcity of sole markings is a function not only of the exposures available, but also of the low velocity of the currents depositing the Back Creek Siltstone.

A third external sedimentary structure is found on the tops of beds of Back Creek Siltstone. Ripple marks are well-exposed on the quarry face at Mouth of Seneca Southwest section, Pendleton County (W-38). These ripple marks can be interpreted to represent the "c" Bouma sequence division.

Paleocurrent data were obtained at four localities: Sugar Camp Run (W-14) (flute casts); Curry Road (W-9)(flute casts); Naval Reservation Headquarters (W-45) (flute casts); and Mouth of Seneca Southwest (W-38) (flute and groove casts).

The best examples of current markings found in the Back Creek Siltstone are in a roadside shale quarry 0.4 mile southwest of Mouth of Seneca, Pendleton County (W-38) on the northwest side of W. Va. Route 28/ U. S. Route 33. The Back Creek Siltstone is well-exposed there with beds almost vertical. Several blocks of siltstone which have excellent examples of flute and groove casts on them have fallen from the quarry face. Although these blocks are not in place now, they are from the Back Creek Siltstone and provide good opportunities for photographing these kinds of sedimentary structures which are commonly associated with turbidity current deposition.

Although the available data are limited, the paleocurrent directions do help to outline the two depositional tongues (see Figure 4 for paleocurrent data plotted on the isopachous map). McIver (1970) did an extensive study of current markings in Upper Devonian rocks in the central Appalachians, with much broader sample spacing then our present study. He was primarily interested in the paleocurrent information and not in the stratigraphic position of the beds for which he determined the direction of depositional flow. However, McIver's (1970) general paleocurrent patterns for the Upper Devonian in the northern part of the Virginias indicate westward-flowing currents, which is consistent with the results obtained in Virginia and West Virginia by Kepferle and others (1978).

## ENVIRONMENT OF DEPOSITION

The Back Creek Siltstone was apparently deposited by turbidity currents moving westward from the eastern margin of the Late Devonian

Appalachian Basin. The earliest evidence for the Augusta Lobe of the Catskill delta is recorded by the Back Creek Siltstone. Since no distinctive siltstones are found at similar stratigraphic positions (Dennison, 1970, 1971; Dennison and Head, 1975) on the other delta lobes (the Fulton, Snyder, and Wyoming Lobes of Willard, 1939), a local origin is implied for the Back Creek Siltstone. A more regional cause such as eustatic sea level drop should have caused similar coarser clastic bundles on the other delta lobes at approximately the same stratigraphic position.

The siltstone beds of the Back Creek Siltstone probably were the result of slumping of some sand with mostly silt and clay off the oversteepened delta front of the Augusta Lobe. This could have created turbid water which then flowed down a hypothetical submarine canyon and out onto the fan and eventually out onto the basin floor. The slope of the submarine fan need not be very steep, nor would the water have to be very deep. Walker and Mutti (1973, p. 119) state that "the ONLY DEPTH CONNOTATION of turbidites is that they be deposited below storm wave base (which is commonly only tens of meters of water)." Modern continental slopes are not very steep; Shepard (1973, p. 279) states that the typical continental slope is not precipitous, averaging 4° 17' for the first 1800 meters of descent.

Probably the Devonian basin margin was not greatly different in slope from a modern continental edge, and had a relatively low slope angle which very slowly became oversteepened as the Augusta Lobe source area was eroded and the Augusta Lobe built seaward. This would cause slumping of coarser detritus off the slope to create turbid water which flowed with relatively low velocity out onto the fan and into the basin to deposit suspended sediment.

The marine origin of the Back Creek Siltstone is established from the general Upper Devonian facies patterns of the units above and below the Back Creek Siltstone horizon. Also, a few marine fossils identifiable only as pelecypods or brachiopods were found in float of the Back Creek Siltstone. The Brallier Formation is generally not very fossiliferous; Butts (1940, p. 319-320) and Woodward (1943, p. 442-443) give short faunal lists for the Brallier Formation in Virginia and West Virginia, respectively. The species diversity is low and the size of the organisms is in general small.

The Back Creek Siltstone is most similar to deposits found on the outer portion of modern and ancient submarine fans described by Ricci-Lucchi (1975, p. 8), Walker and Mutti (1973, p. 142), Nelson (1975, p. 6-1), Huang and Goodell (1970, p. 2071), Mutti and Ricci-Lucchi (1978, p. 150-152) and many others. The two depositional tongues, the internal and external sedimentary structures, the general stratigraphic framework of the Augusta Lobe of the Catskill delta, and the fine-grained nature of the Back Creek Siltstone all provide evidence in favor of the outer fan environment of deposition.

## Evidence For Turbidity Current Origin

The classic concept of a turbidity current deposit sequence was introduced by Bouma (1962) as a group of five divisions, designated "a" through "e". The intervals represent upward-decreasing grain size and current velocities. In the past decade, extensive work on modern and ancient turbidites has revealed that these idealized complete Bouma sequences are rather uncommon in the geologic record as well as in most modern environments.

The Back Creek siltstones and shales are thought to represent the "c", "d", and "e" Bouma divisions. The dominance of the "d" and "e" divisions suggests that these turbidites are distal. Bouma and Hollister (1973, p. 89) state: "Distal turbidites can be defined as turbidity current deposits on the opposite (seaward) side or downslope from the sediment source . . . distal turbidites are totally deposited under low flow regime conditions. If the 'a' and 'b' divisions of Bouma's turbidite facies model are absent, the deposit is then called a distal turbidite (Bouma, 1962, 1964, 1965, 1972a, 1972b; Bouma and Shepard, 1964)."

The association of the "d" and "e" Bouma divisions and some "c" divisions with the internal and external sedimentary structures, the distinct stratigraphic position, and the areal extent of the Back Creek Siltstone are all criteria which support a turbidity current origin.

## Evidence For A Submarine Fan Environment

Turbidites are frequently associated with submarine fan complexes in modern and ancient environments. The submarine fan complex is to be considered as part of the delta complex according to Moore and Asquith (1971, p. 2560) because: " 1) the sediment mass collectively is part of one or several geographically related river deposits; 2) as originally defined or interpreted, the term included, by implication at least, sediments that are offshore from the delta; and 3) recent studies have established that the configuration of the entire system includes the subaerial delta and related proximal deposits, a submarine canyon cut into the slope, and a deep-sea fan with its apex near the mouth of the canyon."

Nelson and Kulm (1973, p. 39) list the requirements for the formation of submarine fans as: 1) a source of sediment; 2) submarine canyons to funnel sediment; and 3) decreasing slope at the lower end of the canyon.

Distal turbidites are found on the lower portion of submarine fans as well as on the relatively flat-lying plains which extend seaward from the edges of the fans. Huang and Goodell (1970, p. 2071) studied the east Mississippi cone in the Gulf of Mexico and made the following statement: "Turbidity currents are believed to occur commonly on deep-sea fans because: 1) a fan owes its existence to a point source of

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sediment derived from upslope, 2) depositional rates on fans are considered to be higher than those on the adjacent slope and rise, and 3) instabilities resulting from rapid accumulation of sediments could lead to slumping, thus initiating flow. "

The association of the distal turbidite facies as well as the presence of the two depositional tongues are criteria for suggesting an outer fan environment of deposition for the Back Creek Siltstone. Ricci-Lucchi (1975, p. 8) describes depositional lobes (or tongues as used here for the detailed configuration of the Back Creek Siltstone associated with the Augusta Lobe of the Devonian delta complex) as areas of thick sand accumulations which form "elongate, prominent bodies aligned with the mouths of the midfan distributaries and passing downward and laterally into thinner deposits without erosional contacts. . . . the lobes are included in the outer fan because of its definition as a non-channeled area. " No channels or suggestions of them have been found in the Back Creek Siltstone, and there are two geographically distinct depositional tongues present.

The absence of channels, the distal nature of the turbidites, and the presence of the two depositional tongues are criteria which can be used to interpret the Back Creek Siltstone as an outer fan deposit.

The presence of the outer fan deposit of Back Creek Siltstone is the only detectable result of the earliest activity on the Augusta Lobe of the Catskill delta which is preserved today. The hypothetical middle and inner fan, the submarine canyon and the subaerial part of the delta which provided sediment for the Back Creek Siltstone are not preserved in the Brallier Formation outcrops observed in the Valley and Ridge Province. These parts of the delta system presumably were located east of the present limits of preservation of Upper Devonian strata in a direction toward the direction of the Blue Ridge Mountains from the area of preserved Back Creek Siltstone occurrence.

### Triggering Mechanisms

A wide variety of mechanisms have been suggested as the causes of slumping and the resulting turbidity currents. Some of the most commonly cited mechanisms as listed by Huang and Goodell (1970, p. 2095) are: "oversteep slopes, rapid sedimentation along the delta front, differential compaction, near-slope current eddies, earthquakes and the decomposition of plant detritus. "

Middleton and Hampton (1973, p. 27) further elaborate on other mechanisms; ". . . in many areas slumping due mainly to rapid deposition of sediment on a slope, leading to underconsolidation of muds, formation of large excess pore fluid pressures, and consequent low internal angles of friction (Morgenstern, 1967). Spontaneous liquefaction of sensitive muds or sands may be produced by relatively minor triggering events. "

The large number of siltstone beds in the Back Creek interval

with considerable thicknesses of intervening shale (representing slower rates of sedimentation, presumably), suggest repeated episodes of slumping and turbidity current flow. Earthquakes would have to have occurred quite frequently to account for all of the siltstone beds in the Back Creek Member, so one of the other mechanisms, such as gradual oversteeping, seems more plausible as a triggering mechanism to initiate Back Creek Siltstone turbidity flows. The time span represented by the Back Creek Siltstone interval is certainly less than a million years (Figure 3), and probably represents a time span of hundreds of thousands of years.

## TECTONIC CONTROLS DURING SEDIMENTATION

The abrupt northern margin of the Back Creek Siltstone and the thickness of the Back Creek at Mouth of Seneca North (W-39) lead us to speculate about the configuration of the basin during deposition. The large thickness of the Back Creek Siltstone at Mouth of Seneca North could be due to piling up or ponding of sediments in a topographic low, similar to the occurrence described by Rupke (1976) in Eocene flysch in the southwestern Pyrenees, and also that described by VanAndel and Komar (1969) in Quaternary sediments in the mid-Atlantic ocean.

The suggestion of some type of syndepositional growth faulting or pronounced syndepositional warping of the basin is supported by two pieces of evidence: 1) the coincidence of the south termination of the Pokejoy Member of the Mahantango Formation with the northern margin of the Back Creek Siltstone (Figure 6) (The Pokejoy Member is characterized by a shallow-water fauna of corals and brachiopods, suggesting an older topographic high north of the Back Creek Siltstone area); and 2) the location of the northern of a pair of photolineaments recognized by Werner (1975, 1976, 1979) on Landsat photographs. The northern lineament (Figure 1) cuts across the Back Creek Siltstone area just south of its northern termination. Werner's photolineament perhaps reflects a growth structure which affected the topography of the Appalachian Basin at the time of deposition of the Back Creek Siltstone.

## ECONOMIC POTENTIAL

The Back Creek Siltstone has been identified in three anticlinal gas fields, drilled to the Oriskany Sandstone target a few hundred feet below the position of the Back Creek Siltstone. The fields are the Thornwood and Glady fields in West Virginia and the Bergton field in Rockingham County, Virginia (Figure 13). No gas has ever been reported from the Back Creek Siltstone to the writers' knowledge; however, the Back Creek Siltstone could have potential as a reservoir.

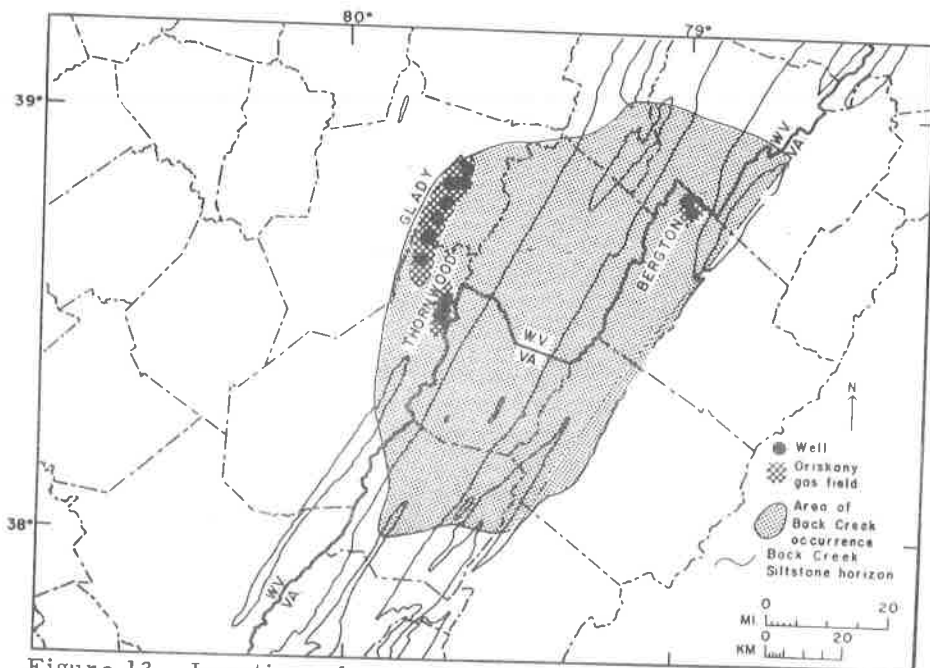The fractures, many of them open, and the environment of

Figure 13.   Location of anticlinal Oriskany Sandstone gas fields
underlying the Back Creek Siltstone.

deposition make the  Back  Creek a possible reservoir.    The outer fan
environment is  located adjacent to the basin plain pelagic shales which
are likely source  rocks.  Nelson (1975, p.  6-4) states:  "Depositional
lobes of outer fans  may  be traps for petroleum accumulation because
of their wide lateral extent  and  the  great  quantities  of  hemipelagic
source beds of basin  plain  deposits  that  enclose and possibly help to
seal them. "

Glaeser (1978, p.  517) believes  that  the  Devonian slope sedi-
ments in the central Appalachian Basin have economic potential, as he
states:  "Clastic wedge deposits in the regressive sequence built by the
Catskill delta  include  accumulations  of  terrigenous-derived offshore
source-beds and nearshore reservoir-bed silts and sands. "

Certainly, the stratigraphic position of the Back Creek Siltstone
above the  Middle Devonian dark shales,  its well-developed fractures,
and  its  environment  of  deposition  all  are favorable attributes for a
methane reservoir. The  small  grain  size of Back Creek detritus and
the extremely fine  matrix are unfavorable for gas reservoir develop-
ment, but the eastern coarser areas have more interesting attributes.

Several similar zones  such  as  the Benson sandstone and silt-
stone (Cheema  and  others,  1977) and the siltstones of the Sycamore
"grit" (Nock and Patchen, 1975) are active gas-producing stratigraphic
units at the  present  time  in  central West Virginia.   These producing

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units are somewhat higher in the Devonian stratigraphic column and farther west than the Back Creek Siltstones documented in the present study. The Back Creek Siltstone represents a more easterly and somewhat older occurrence of a similar potential drilling target, and merits consideration for methane exploration.

## REFERENCES CITED

Avary, K. L., 1978, Stratigraphy and sedimentology of the Back Creek Siltstone (Devonian) in Virginia and West Virginia: Unpub. M. S. thesis, Univ. of North Carolina, Chapel Hill, N. C., 89 p.

Bennetts, K. R. W., 1974, Characteristics of three individual turbidites from the Hispaniola-Caicos Basin: Unpub. M. S. thesis, Duke Univ., Durham, North Carolina, 131 p.

Bouma, A. H., 1962, Sedimentology of some flysch deposits: Amsterdam, Elsevier, 168 p.

_____, 1964, Turbidites, in Bouma, A. H. and Brouwer, A., eds., Turbidites: Developments in sedimentology 3, Amsterdam, Elsevier, p. 247-256.

_____, 1965, Sedimentary characteristics of samples collected from some submarine canyons: Marine Geology, v. 3, p. 291-320.

_____, 1972a, Distribution of sediments and sedimentary structures in the Gulf of Mexico: in Rezak, R. and Henry, V. J., eds., Contributions on the geological and geophysical oceanography of the Gulf of Mexico: Texas A & M Oceanographic Studies: Houston, Gulf Publishing Co., v. 3, p. 35-65.

_____, 1972b, Recent and ancient turbidites and contourites: Gulf Coast Assoc. Geol. Socs. Trans., v. 22, p. 205-221.

Bouma, A. H., and Hollister, C. D., 1973, Deep ocean basin sedimentation, in Middleton, G. V. and Bouma, A. H., eds., Turbidites and deep-water sedimentation: Anaheim, Soc. Econ. Paleontologists and Mineralogists, Pacific Section, Short Course Notes, p. 79-118.

Bouma, A. H., and Shepard, F. P., 1964, Large rectangular cores from submarine canyons and fan valleys: Am. Assoc. Petroleum Geologists Bull., v. 48, p. 225-231.

Butts, C., 1918, Geologic section of Blair and Huntingdon Counties, central Pennsylvania: Am. Jour. Sci., 4th ser., v. 46, p. 523-537.

_____, 1933, Geologic map of the Appalachian Valley of Virginia with explanatory text: Virginia Geol. Survey Bull. 42, 56 p.

_____, 1940, Geology of the Appalachian Valley in Virginia: Virginia Geol. Survey Bull. 52, Part I, Geologic text and illustrations, 568 p.

Cheema, M. B., Donaldson, A. C., Heald, M. T., and Renton, J. J.,

, 1977, Gas-producing submarine fan deposits of Catskill delta complex in north-central West Virginia: (abs.) Am. Assoc. Petroleum Geologists Bull., v. 61, p. 775-776.

Dennison, J. M., 1960, Stratigraphy of Devonian Onesquethaw Stage in West Virginia, Virginia, and Maryland: Unpub. Ph. D. dissertation, Univ. of Wisconsin, Madison, Wis., 339 p.

_____, 1961, Stratigraphy of Onesquethaw Stage of Devonian in West Virginia and bordering states: West Virginia Geol. Survey Bull. 22, 87 p.

_____, 1965, Devonian clastics along Allegheny Front from Allegany County, Maryland to Highland County, Virginia: (abs.) Geol. Soc. America Spec. Paper 82, p. 297.

_____, 1970, Stratigraphic divisions of Upper Devonian Greenland Gap Group ("Chemung Formation") along Allegheny Front in West Virginia, Maryland, and Highland County, Virginia: Southeastern Geology, v. 12, p. 53-82.

_____, 1971, Petroleum related to Middle and Upper Devonian deltaic facies in central Appalachians: Am. Assoc. Petroleum Geologists Bull., v. 55, p. 1179-1193.

Dennison, J. M., and DeWitt, W., Jr., 1972, Redbed zone produced by sea level drop at beginning of Devonian Cohocton Age delimits Fulton and Augusta lobes at Catskill delta complex, in Guidebook 37th Ann. Field Conf. of Pennsylvania Geologists: Harrisburg, Pennsylvania, Pennsylvania Topog. and Geol. Survey, p. 109-114.

Dennison, J. M., and Hasson, K. O., 1976, Stratigraphic cross section of Hamilton Group (Devonian) and adjacent strata along south border of Pennsylvania: Am. Assoc. Petroleum Geologists Bull., v. 60, p. 278-287.

_____, 1979, Stratigraphic cross section of Devonian shales along the Allegheny Front from Maryland to Highland County, Virginia: West Virginia Acad. Sci. Proc., v. 49, p. 103-110.

Dennison, J. M. and Head, J. W., 1975, Sealevel variations interpreted from the Appalachian basin Silurian and Devonian: Am. Jour. Sci., v. 275, p. 1089-1120.

Dennison, J. M. and Naegele, O. D., 1963, Structure of Devonian strata along Allegheny Front from Corriganville, Maryland to Spruce Knob, West Virginia: West Virginia Geol. and Economic Survey Bull. 24, 42 p.

Enos, P., 1969, Anatomy of a flysch: Jour. Sedimentary Petrology, v. 39, p. 680-723.

Folk, R. L., 1974, Petrology of sedimentary rocks: Austin, Texas, Hemphill Publishing Co., 182 p.

Frakes, L. A., 1967, Stratigraphy of the Devonian Trimmers Rock in eastern Pennsylvania: Pennsylvania Topographic and Geol. Survey General Geology Report G 51, 4th ser., 208 p.

Glaeser, J. D., 1978, Devonian slope sediments in Central Appalachian Basin-source and reservoir deposits: (abs.) Am. Assoc. Petroleum Geologists Bull., v. 62, p. 517.

Harland, W. B., Smith, A. G., and Wilcox, B., eds., 1964, The Phanerozoic time-scale; a symposium dedicated to Professor Arthur Holmes: Quart. Jour. Geol. Soc. London, v. 120s, 458 p.

Hasson, K. O. and Dennison, J. M., 1978, Devonian Harrell and Millboro Shales in parts of Pennsylvania, Maryland, West Virginia, and Virginia: in G. L. Schott, W. K. Overby, Jr., A. E. Hunt, and C. A. Komar, eds., Proceedings First Eastern Gas Shales Symposium, October 17-19, 1977, Morgantown, West Virginia: U. S. Department of Energy, Morgantown Energy Research Center, Publication MERC/SP-77/5, p. 634-641.

Heckel, P. H., 1969, Devonian Tully Limestone in Pennsylvania and comparison to type Tully Limestone in New York: Pennsylvania Topographic and Geol.

Surv., 4th ser., Inf. Circ. 60, 33 p.

Huang, Ter-Chien and Goodell, H. G., 1970, Sediments and sedimentary process-
es of eastern Mississippi cone, Gulf of Mexico: Am. Assoc. Petroleum
Geologists Bull., v. 54, p. 2070-2100.

Ingram, R. L., 1954, Terminology for the thickness of stratification and parting
units in sedimentary rocks: Geol. Soc. America Bull., v. 65, p. 937-938.

Kepferle, R. C., Lundegard, Paul, Maynard, J. B., Potter, P. E., Pryor, W. A.,
Samuels, Neil, and Schauf, F. J., 1978, Paleocurrent systems in shaly
basins: preliminary results for Appalachian basin (Upper Devonian): in
G. L. Schott, W. K. Overby, Jr., A. E. Hunt, and C. A. Komar, eds.,
Proceedings First Eastern Gas Shales Symposium, October 17-19, 1977,
Morgantown, West Virginia: U. S. Department of Energy, Morgantown
Energy Research Center, Publication MERC/SP-77/5, p. 434-441.

Kuenen, P. H., 1966, Matrix of turbidites: experimental approach: Sedimentology,
v. 7, p. 267-297.

Kulander, B. R., 1968, A structural analysis of Browns Mountain anticline in West
Virginia: Unpub. Ph. D. dissert., West Virginia Univ., Morgantown, 174 p.

Lundegard, P. D., Samuels, N. D., and Pryor, W. A., 1978, The Brallier For-
mation- Upper Devonian turbidite slope facies of the central and southern
Appalachians, in Preprints, Second Eastern Gas Shales Symposium, October
16-18, 1978, Morgantown, West Virginia: U. S. Dept. Energy, Morgan-
town Energy Technology Center, Publication METC/SP-78/6, vol. 1, p. 4-
23.

McBride, E. F., 1962, Flysch and associated beds of the Martinsburg Formation
(Ordovician), Central Appalachians: Jour. Sed. Petrology, v. 32, p. 39-91.

McGhee, G. R., Jr. and Dennison, J. M., 1976, The Red Lick Member, a new
subdivision of the Foreknobs Formation (Upper Devonian) in Virginia, West
Virginia, and Maryland: Southeastern Geology, v. 18, p. 49-57.

McIver, N. L., 1970, Appalachian turbidites, in Fisher, G. W., Pettijohn, F. J.,
Reed, J. C., Jr., and Weaver, K. N., eds., Studies of Appalachian Geo-
logy: Central and Southern: New York, Interscience, p. 69-81.

Middleton, G. V. and Hampton, M. A., 1973, Sediment gravity flows: mechanics
of flow and deposition, in Middleton, G. V. and Bouma, A. H., eds., Tur-
bidites and deep-water sedimentation: Anaheim, Soc. Econ. Paleontologists
and Mineralogists, Pacific Section, Short Course Notes, p. 1-38.

Moore, G. R. and Asquith, D. O., 1971, Delta: term and concept: Geol. Soc.
America Bull., v. 82, p. 2563-2567.

Morgenstern, N. R., 1967, Submarine slumping and the initiation of turbidity cur-
rents, in Richards, A. F., ed., Marine Geotechnique: Urbana, Univ. of
Illinois Press, p. 189-220.

Mutti, E. and Ricci-Lucchi, 1978, Turbidites of the Northern Appenines: Intro-
duction to facies analysis: International Geology Review, v. 20, p. 125-166
(Am. Geol. Institute Reprint Series 3).

Nelson, C. H., 1975, Turbidite fans and other base-of-slope deposits, in Dickin-
son, W. R., ed., Current concepts of depositional systems with applications
for petroleum geology: San Joaquin Geological Society Short Course Notes,
p. 6-1 - 6-5.

Nelson, C. H., and Kulm, V. D., 1973, Submarine fans and deep-sea channels, in
Middleton, G. V. and Bouma, A. H., eds., Turbidites and deep-water sedi-
mentation: Anaheim, Soc. Econ. Paleontologists and Mineralogists Pacific
Section, Short Course Notes, p. 39-78.

Nock, H. M. and Patchen, D. G., 1975, Stratigraphy of the Sycamore siltstone
(Upper Devonian) in West Virginia: Geol. Soc. of America, Abstracts with
Programs (Northeastern Section), v. 7, no. 1, p. 101.

Oliver, W. A., Jr., DeWitt, W., Jr., Dennison, J. M., Hoskins, D. M., and

Huddle, J. W., 1967, Devonian of the Appalachian basin, United States, in Oswald, D. H., ed., International symposium on the Devonian System: Alberta Soc. Petroleum Geologists, v. 1, p. 1001-1040.

Oliver, W. A., Jr., DeWitt, W., Jr., Dennison, J. M., Hoskins, D. M., and Huddle, J. W., 1969, Correlation of Devonian rock units in the Appalachian Basin: U. S. Geol. Survey Oil and Gas Inves. Chart OC-64.

Pettijohn, F. J. and Potter, P. E., 1964, Atlas and glossary of primary sedimentary structures: New York, Springer-Verlag, 370 p.

Picard, M. D., 1971, Classification of fine-grained sedimentary rocks: Jour. Sed. Petrology, v. 41, p. 179-195.

Ricci-Lucci, F., 1975, Depositional cycles in two turbidite formations of northern Apennines (Italy): Jour. Sed. Petrology, v. 45, p. 3-43.

Rickard, L. V., 1975, Correlation of the Silurian and Devonian rocks in New York State: New York State Mus. and Sci. Service, Geol. Survey, Map and Chart Ser. No. 24.

Rupke, N. A., 1976, Sedimentology of very thick calcarenite-marlstone beds in a flysch succession, southwestern Pyrenees: Sedimentology, v. 23, p. 43-65.

Schenk, P. E., 1970, Regional variation of the flysch-like Meguma Group (Lower Paleozoic) of Nova Scotia, compared to recent sedimentation off the Scotian Shelf, in Lajoie, J., ed., Flysch sedimentology in North America: Geol. Assoc. Canada Spec. Paper Number 7, p. 127-154.

Shepard, F. P., 1973, Submarine geology, 3rd edition: New York, Harper and Row, 517 p.

Tilton, J. L., Prouty, W. F., and Price, P. H., 1927, Pendleton County, West Virginia Geol. Survey, 384 p.

VanAndel, T. H. and Komar, P. D., 1969, Ponded sediments of the mid-Atlantic ridge between $22^\circ$ and $23^\circ$ north latitude: Geol. Soc. America Bull., v. 80, p. 1163-1190.

Walker, R. G. and Mutti, E., 1973, Turbidite facies and facies associations: in Middleton, G. V. and Bouma, A. H., eds., Turbidites and deep-water sedimentation: Anaheim, Soc. Econ. Paleontologists and Mineralogists, Pacific Section, Short Course Notes, p. 119-157.

Werner, E., 1975, Long lineaments of the Appalachians and their geologic relations: Geol. Soc. America, Abs. with Programs (Southeastern Sec.), v. 7, no. 4, p. 548.

_____, 1976, Photolineaments derived from Landsat imagery related to structural map and field data from southern West Virginia: Geol. Soc. America, Abs. with Programs (Southeastern Sec.), v. 8, no. 2, p. 297-298.

_____, 1979, The 38th Parallel lineament in West Virginia, in Podwysocki, M. H. and Earle, J. L., eds., Proc. Second International Conf. on Basement Tectonics, July 13-17, 1976, p. 406-412.

Willard, B., 1939, Catskill continental facies, in The Devonian of Pennsylvania: Pennsylvania Geol. Survey Bull. G19, p. 257-307.

Woodward, H. P., 1932, Geology and mineral resources of the Roanoke area, Virginia: Virginia Geol. Survey Bull. 34, 172 p.

_____, 1943, Devonian system of West Virginia: West Virginia Geol. Survey, v. 15, 655 p.

Data point (V-16). 38° 27' 6" N, 79° 39' 0" W. Highland County, Virginia. Hightown 7. 5' quadrangle (NE), in road cuts and along farm road adjacent to U. S. Route 250, on the northwest flank of Lantz Mountain, immediately east of the valley of Back Creek (Figure 14). Data from K. L. Avary, J. M. Dennison, and K. O. Hasson.

|  | ft. | m |
|---|---|---|

**Brallier Formation (187+ ft; 57.0 m)**

35.  Not measured, shales and thin siltstones.
34.  Covered, along bend in highway. Prominent siltstones should crop out if present.
33.  Back Creek Siltstone Member (68. 1 ft; 20. 8 m)

|  | | |
|---|---|---|
| Siltstone, light olive gray, weathers moderate yellowish brown; interbedded with shale, medium dark gray, weathers light olive gray. | 68. 1 | 22. 3 |

Siltstone; bed 37.  (20 cm)
Shale, chippy.  (40 cm)
Siltstone; bed 36.  (31 cm)
Shale.  (16 cm)
Siltstone; bed 35.  (9 cm)
Shale.  (22 cm)
Siltstone; bed 34.  (6 cm)
Shale, slightly silty, chippy.  (18 cm)
Siltstone; bed 33.  (18 cm)
Covered.  (360 cm)
Siltstone; bed 32.  (72 cm)
Shale, very silty.  (35 cm)
Siltstone; bed 31.  (19 cm)
Shale, silty, chippy near base.  (50 cm)
Siltstone; shaly at top; bed 30.  (23 cm)
Shale, silty.  (6 cm)
Siltstone; bed 29.  (14 cm)
Shale, weathered.  (5 cm)
Siltstone, shaly at top; bed 28.  (9 cm)
Shale, somewhat silty, chippy.  (22 cm)
Siltstone; bed 27.  (3 cm)
Shale, chippy.  (12 cm)
Siltstone, shaly at top; bed 26.  (43 cm)
Shale, silty, weathered.  (30 cm)
Siltstone; bed 25.  (40 cm)
Shale, mostly chippy.  (27 cm)
Siltstone; bed 24.  (14 cm)
Shale.  (6 cm)
Siltstone, shaly at top; bed 23.  (8 cm)
Shale, chippy.  (6 cm)
Siltstone; bed 22.  (9 cm)
Shale, silty in part, most weathers chippy.  (277 cm)
Siltstone; bed 21.  (30 cm)
Shale, silty at base, top weathers chippy.  (12 cm)
Siltstone; bed 20.  (33 cm)
Shale, very silty at top and base.  (25 cm)
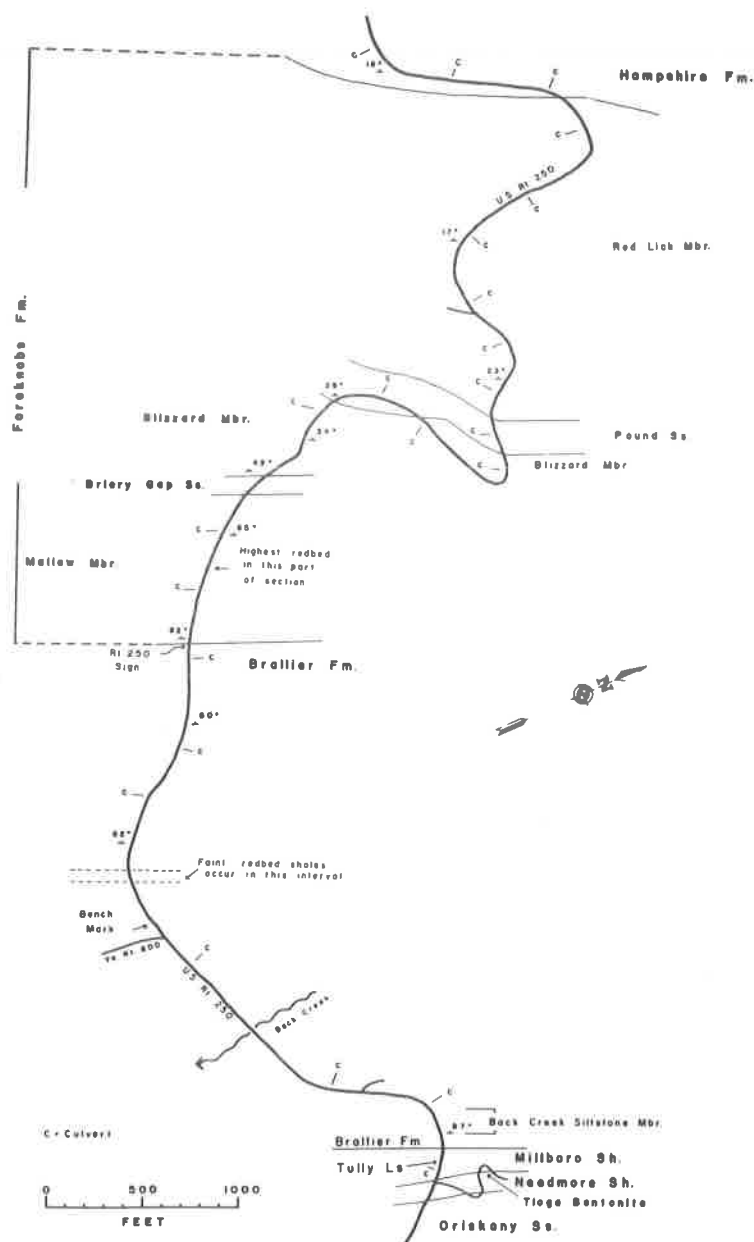Siltstone; bed 19.  (26 cm)

Figure 14. Map of type section of Back Creek Siltstone Member of Brallier Formation. Data point V-16. Highland County, Virginia. Hightown 7.5 Minute Quadrangle. Lat. 38°27'06" N, Long. 79°39'00"W.

Shale, silty at top, weathers chippy.  (9 cm)
Siltstone; bed 18.  (15 cm)
Shale, silty near base, weathers chippy. (32 cm)
Siltstone; bed 17.  (6 cm)
Shale, silty, evenly laminated.  (55 cm)
Siltstone, shaly at top; bed 16.  (38 cm)
Shale.  (17 cm)
Siltstone, shaly at top 2 cm; bed 15.  (58 cm)
Shale.  (1 cm)
Siltstone; bed 14.  (5 cm)
Shale.  (1 cm)
Siltstone; bed 13.  (4 cm)
Shale, chippy.  (2 cm)
Siltstone; bed 12.  (49 cm)
Shale.  (1 cm)
Siltstone; bed 11.  (110 cm)  (This is thickest siltstone
    bed in this section of Back Creek Member).
Shale.  (2 cm)
Siltstone; bed 10.  (9 cm)
Shale.  (3 cm)
Siltstone; bed 9.  (3 cm)
Shale.  (14 cm)
Siltstone, shaly at top; bed 8.  (8 cm)
Shale.  (3 cm)
Siltstone; bed 7.  (10 cm)
Shale.  (5 cm)
Siltstone, shaly at top; bed 6.  (53 cm)
Shale.  (4 cm)
Siltstone; bed 5.  (18 cm)
Shale, thinly laminated, chippy.  (9 cm)
Siltstone; bed 4.  (30 cm)
Shale, chippy.  (12 cm)
Siltstone; bed 3.  (17 cm)
Shale, chippy.  (11 cm)
Siltstone; bed 2.  (16 cm)
Shale, silty in middle, top and base weather chippy.
    (54 cm)
Siltstone; bed 1.  (18 cm) (This is lowest siltstone over
    10 cm thick, which is the arbitrary cutoff for silt-
    stones assigned to the Back Creek).

32. Shales with some thin siltstones. Shale weathers light olive
gray. Siltstones are less than 10 cm thick and weather light
olive gray to moderate yellowish brown.                        119.0   36.3

Millboro Shale (228 ft; 69.5 m)

31. Shale, dark gray, thickly to thinly laminated, slightly silty
at top, weathers light olive gray and chippy.                 32.0   10.5
30. Shale, calcitic, medium dark gray, thickly laminated,
weathers light olive gray and chippy to platy; contains lime-
stone concretions up to 0.3 ft (0.09 m) thick, with three
such concretions exposed in road cut.  This concretion zone
represents the Tully Limestone.                               7.0    2.1

29. Shale, grayish black to very dark gray, thinly laminated weathers yellowish gray and platy. — 74.0 — 22.6
28. Shale, weathers yellowish gray and platy to chippy. Probably very weathered Purcell Member with any limestones dissolved by weathering. — 7.0 — 2.1
27. Covered. — 20.0 — 6.1
26. Shale, dark gray, thinly laminated, weathers platy. — 46.0 — 14.0
25. Shale with abundant tiny fossils, weathers brownish gray. Uppermost Tioga Bentonite tuffaceous unit. — 1.8 — 0.6
24. Shale, grayish black, thinly laminated. — 10.1 — 3.1
23. Shale, with tiny fossils, thinly to thickly laminated, weathers brownish gray and platy to chippy. Tioga. — 12.0 — 3.7
22. Covered. — 6.0 — 1.8
21. Shale, grayish black, thinly laminated, weathers yellowish gray and platy. — 11.0 — 3.4
20. Shale, interbedded brownish black and grayish black, thinly laminated. Tioga tuffaceous interlayers. — 5.0 — 1.5
19. Shale, brownish black, thinly to thickly laminated, weathers brownish gray and platy to chippy; contains abundant tiny fossils characteristic of Tioga Bentonite. Tioga middle coarse zone is probably at base of this interval, but outcrop is too weathered to permit recognition of the sand-size tuff beds. — 13.7 — 4.2

Needmore Shale (106 ft; 32.3 m)

18. Shale, thickly laminated, medium dark gray. — 2.1 — 0.6
17. Shale, thinly laminated, brownish gray. Definite Tioga Bentonite tuffaceous unit. — 1.6 — 0.5
16. Shale, thickly laminated, weathers light olive gray. Definite Needmore Shale. — 3.1 — 1.0
15. Shale, thinly laminated, weathers brownish gray and platy. Lowest definite Tioga Bentonite tuffaceous unit. — 2.0 — 0.6
14. Clay, yellowish gray. Ground water leaching effect? — 0.8 — 0.2
13. Shale, medium dark gray, thickly laminated, weathers light olive gray and chippy. — 13.0 — 4.0
12. Shale, dark to very dark gray, thinly to thickly laminated, weathers medium dark gray and platy to chippy. — 6.8 — 2.1
11. Limestone, argillaceous, very weathered to yellowish gray. — 0.2 — 0.06
10. Shale, dark to very dark gray, thinly to thickly laminated, weathers medium dark gray and chippy to platy. — 10.8 — 3.3
9. Limestone, shaly, very weathered. — 0.2 — 0.06
8. Shale, dark to very dark gray, thinly to thickly laminated, weathers medium dark gray and platy to chippy. — 7.3 — 2.2
7. Limestone, very weathered to yellowish gray. Probably lowest limestone in Needmore Shale. — 0.2 — 0.06
6. Shale, dark gray, thickly laminated, weathers medium dark gray and chippy. — 3.8 — 1.2
5. Siliceous shale, lenticular. — 0.2 — 0.06
4. Shale, dark gray, thickly laminated, weathers medium dark gray and chippy. — 9.0 — 2.7
3. Shale, siliceous, medium dark gray when fresh, thickly laminated to thinly bedded, weathers knobby. Intertongue from west of Huntersville Chert facies. — 11.0 — 3.4

|  | ft | m |
|---|---|---|
| 2. Shale, thinly laminated, weathers light olive gray to olive gray. Beaver Dam Member of Needmore Shale. | 29.6 | 9.0 |
| 1. Siltstone, sandy, thickly laminated, weathers light olive gray. | 0.4 | 0.1 |

Oriskany Sandstone (not measured)

# TOPOGRAPHIC MAPPING OF SHELF EDGE PROMINENCES OFF SOUTHEASTERN FLORIDA[1]

By

M. John Thompson[2]
Lewis E. Gilliland
Remote Sensing Services, Harbor Branch Foundation, Inc.
Ft. Pierce, Florida 33450

## ABSTRACT

A zone of topographic prominences along Florida's eastern continental shelf edge has been mapped, using side-scan sonar and fathometer tracings. This zone (27°49.5N and 79°58.5'W), designated "Sebastian Pinnacle System", is extremely complex and contains a concentration of major shelf edge features resulting from combined geophysical and biological forces. Substrate distribution patterns indicate strong depositional and erosional activity by the Florida Current. Numerous holes and crater-like depressions located throughout the area reflect differential erosion and/or sapping of the underlying limestone during periods of lower sea level. A series of submersible dives were made to investigate specific bottom features seen on the sonographs and fathometer tracings. These dives confirmed the complexity of the overall system, and observations made at several key locations indicate the major relief results from two types of drowned topographic features, mounds of oolitic limestone and relict coral reefs.

## INTRODUCTION

The existence of numerous topographic prominences along the continental margins of southeastern Florida has been well documented (Uchupi, 1966 and 1969; Macintyre and Milliman, 1970; Avent et al., 1977). Recent work done in this area (Avent et al., 1977) presents a general description of morphology and associated biota for the area between Cape Canaveral and West Palm Beach. To date, there are no published studies detailing the topography of such shelf edge features. During 1977, precision depth recorders and side-scan sonar were used to study the detailed topography of one area showing an unusual concentration of prominences (Figure 1). This area has been designated the "Sebastian Pinnacle System" and centers about Lat. 27°49.5'N and Long. 79°58.5'W.

[1] Contribution No. 160 of the Harbor Branch Foundation, Inc.
[2] Current adress: All Points Realty, 1116 NW Fed. Hwy., Stuart, FL 33494
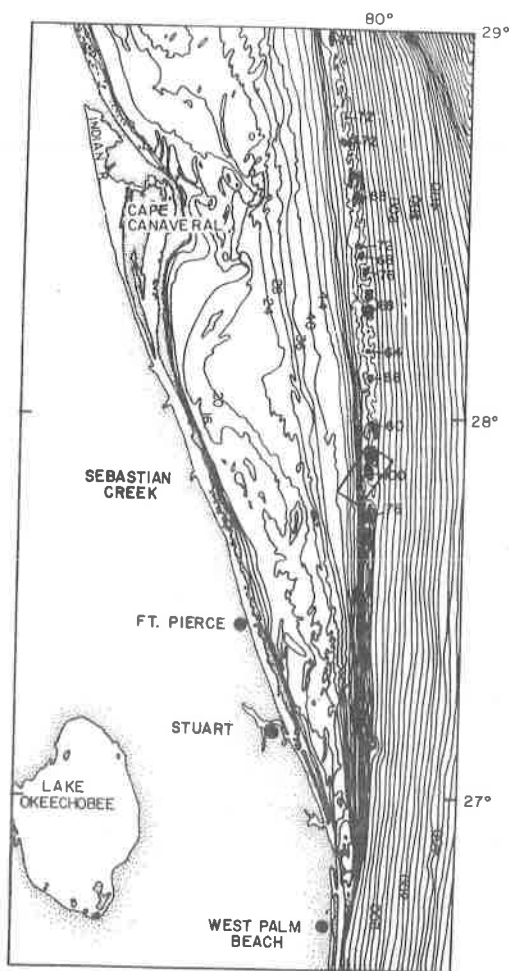
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Figure 1.  Shelf edge prominences off south-
eastern Florida.  The "Sebastian
Pinnacle System" is contained with-
in the designated rectangular area.
(This figure is modified from
Uchupi, 1966).

## METHODS

Preliminary surveys of the Sebastian Pinnacle System were con-
ducted during the spring and summer of 1977 from R/V GOSNOLD.
From these missions, a study area bounded by the following coordinates
was selected; northeast corner - 27°53'N 79°54.5'W, northwest corner -
27°53. 5'N 79°58. 5'W, southeast corner - 27°45. 5'N 79°58'W, south-
west corner - 27°46. 5'N 80°02'W (Figure 1).  Bathymetric mapping

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within this study area was completed in November of 1977, from the survey vessel SEA DIVER.

During field operations, navigational fixes were taken using a Northstar[R] 6000 automatic LORAN C receiver. Accuracy of the LORAN C network used in this study area was ±850 m (0.46 nmi) north to south. Depth profile data was obtained using a Raytheon[R] DE 121 fathometer (Thompson et al., 1978).

The Klein[R] Series 400 Side-Scan Sonar System used in this study produced a view of seafloor shape and texture to a maximum range of 150 meters on either side of the ship. Calibration of sonograph tonal quality was conducted in a large sand flat area (depth 25-50 m), following the recommendations of Flemming (1976).

Five areas within the Sebastian Pinnacle System were chosen for observation using JOHNSON-SEA-LINK II (J-S-L II) submersible. These areas are identified in Table 1 and indicated by the broken line boxes on the included maps (Figure 2; A, B, and C). They were specifically selected because each contains characteristic features seen throughout the Sebastian Pinnacle System.

## RESULTS AND DISCUSSION

Topographic maps contained here are presented on LORAN C navigational grids. Because of the size of the Sebastian Pinnacle study area (110 km$^2$), its topography is presented in three overlapping charts (Figure 2; A, B, and C).

Systematic topographic mapping verified the extreme complexity of the Sebastian Pinnacle area. This system contains bifurcating ridge complexes, irregular knolls, cones, multi-peaked hills and crater like depressions. Substrate composition is equally complex. Rocks, shell hash, coral debris and sand characterize substrata seen in a mosaic of rocky prominences, exposed pavement, ledges, talus accumulations and unconsolidated debris. Living Oculina varicosa was reported as scarce in this area by Avent et al. (1977) and this observation was confirmed by our submersible dives. However, Oculina rubble is very abundant on the southeastern slopes and crests of the pinnacles (Figure 3).

All major prominences were characterized by depressions or troughs at their northern and/or southern bases. The bottoms of these depressions appeared to be subject to considerable scouring action, and in some cases basal pavements were exposed (Figure 4).

While maneuvering through one of the valleys at the High Alpha site, a large (10 m tall by 20 m long) projection of consolidated limestone, roughly pyramidal in shape, was encountered which apparently had been exposed by the action of currents within the valley. Many surface blemishes were visible of the type characteristically associated with solution features in weathered rock facies.
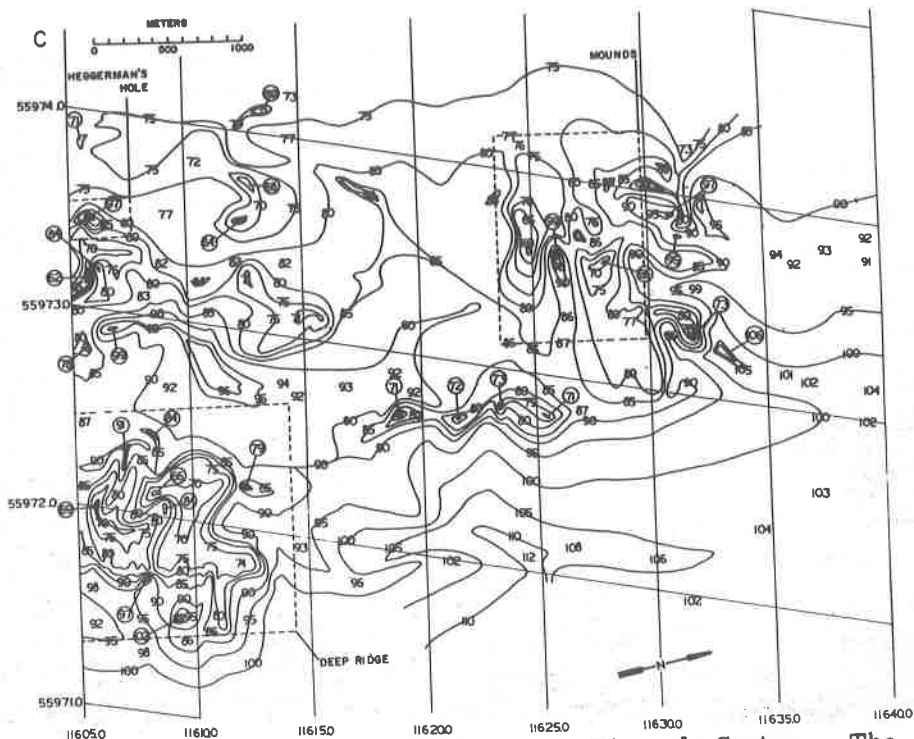
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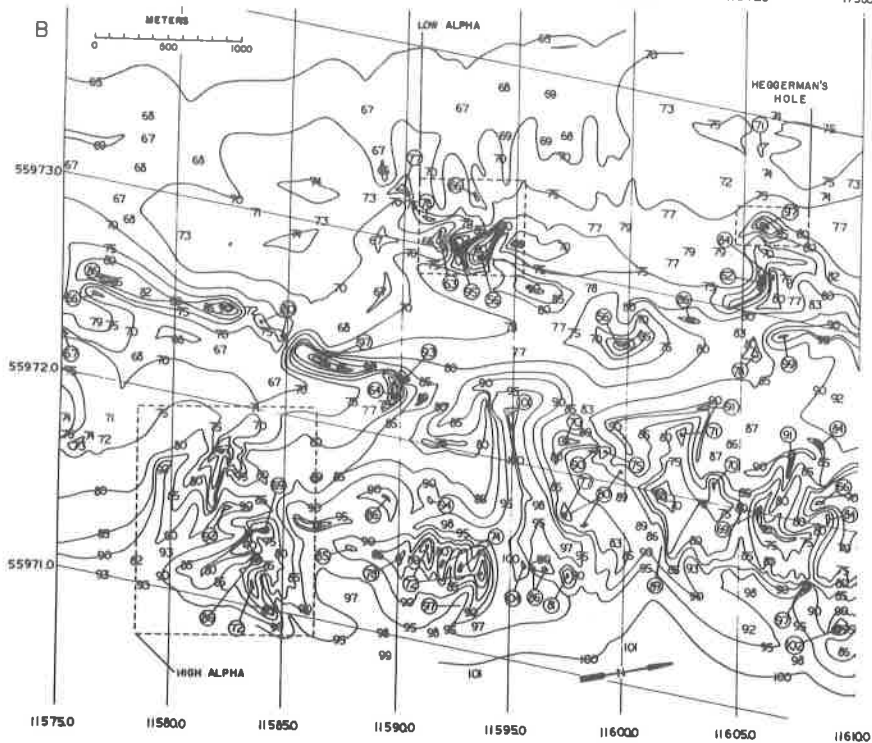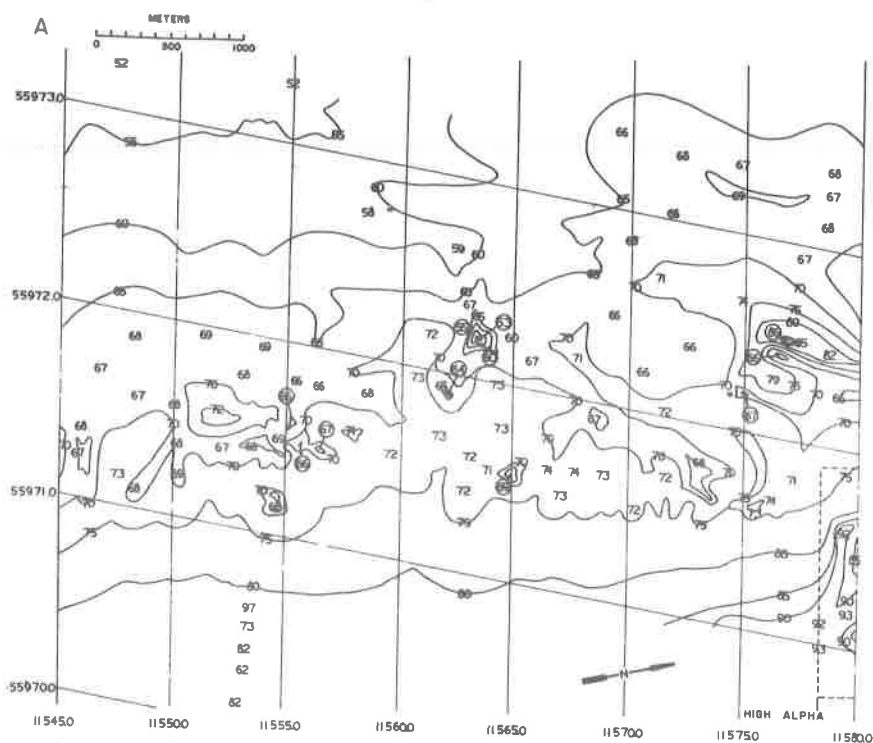
Figure 2.  Contour map of the Sebastian Pinnacle System.  The contour interval is 5 meters and all depths indicated are corrected to mean low tide.
   A.  Southern section, 11545.0 to 11580.0 microseconds (LORAN C)
   B.  Middle section, 11575.0 to 11610.0 microseconds (LORAN C)
   C.  Northern section, 11605.0 to 11640.0 microseconds (LORAN C)

Table 1.  Sites within the Sebastian Pinnacle System where observation dives were made with JOHNSON-SEA-LINK II.

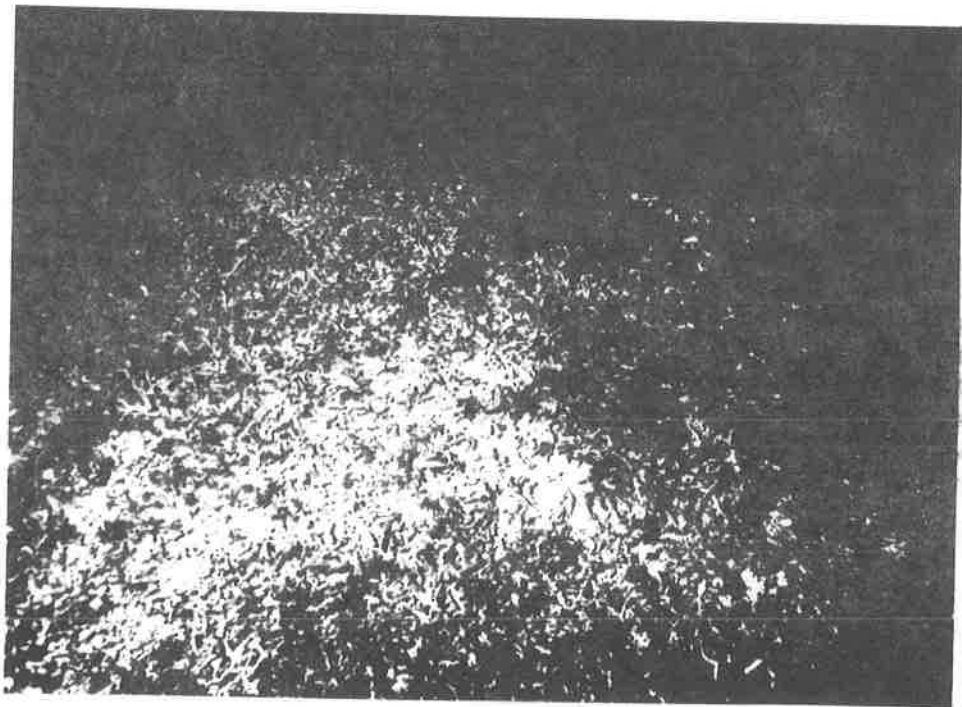| Designation | LORAN C Coordinates | North Lat. | West Long. |
|---|---|---|---|
| High Alpha | 11582.5 - 55971.4 | 27°48.7' | 79°57.6' |
| Deep Ridge | 11607.5 - 55972.0 | 27°50.5' | 79°57.3' |
| Low Alpha | 11593.0 - 55973.2 | 27°49.7' | 79°58.7' |
| Mounds | 11627.0 - 55973.7 | 27°52.3' | 79°57.9' |
| Heggerman's Hole | 11606.0 - 55973.4 | 27°50.6' | 79°58.5' |

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Figure 3. Oculina varicosa rubble on southeastern slope of a ridge.

The Deep Ridge study area showed ridges with very sharp crests. These ridges were oriented perpendicular to the shelf edge break. Overall structure in this area is reminiscent of an exaggerated spur and groove system. Ridge crests gradually slope down to a flat plain in approximately 110 meters of water. Observation dives again showed extensive unconsolidated Oculina rubble but very few living colonies. Very rapid, current-driven sediment transport over the ridge crests was observed frequently in this area. A thick layer of accumulated debris and sediment obscured the underlying structure of the crest zone. Further down the ridge slope, large coral heads, dead but in place, of what appeared to be Montastrea annularis were observed. Macintyre and Milliman (1970) report collecting four species of hermatypic corals, including M. annularis, from dredge samples taken in this same general area.

The extreme topographic variation seen throughout the Sebastian Pinnacle System produces numerous subsurface gyres and eddys in the Florida Current as it flows along this shelf edge feature. Bottom currents were extremely strong and variable throughout the entire study area. Currents between 1.0 and 1.5 knots were commonly encountered and investigations at the Low Alpha site were not completed because heavy turbulence interfered with submersible operation.

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Figure 4.   Exposed basal pavements in submerged valley.

The Mounds, located at the northern end of the study grid (Figure 2, C) proved to be one large dome-like structure with several smaller ridges near it.  These ridges were low and had rounded summits in contrast to those of the Deep Ridge area. Shell hash and coarse rubble dominated the sediments on the southern and western sides of these structures, grading into sand on the northern slopes.

Heggerman's Hole is one of several large crater-like depressions seen in the Sebastian Pinnacle System.  This particular depression drops from a depth of 77 m, down to 100 m, with a slope ranging from $10^{\circ}$ to $20^{\circ}$.  Although there are some small ridges on the north side of the hole, there is nothing which could be considered a major mound or pinnacle near this depression.  The bottom of Heggerman's Hole was scoured clean and showed a limestone-like basal pavement. Some rock boulders were seen on the northern side and a rock slope was encountered on the southwestern side.  It is possible that this feature, and the other craters seen, may be solution features or sinkholes formed by the dissolving of softer limestone sections of the shelf edge ridge in this area.  These are similar to other submerged depressions seen south of the Florida Keys and in the Bahama Banks (Uchupi, 1968).

It is generally believed that the shelf edge prominences seen along the southeastern coast of Florida resulted from several geological

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processes. The underlying structure was formed by reef building during the Tertiary, but this structure was vastly modified by subsequent deposition during the Pleistocene and, in recent times, erosion by the Florida Current (Uchupi, 1969). The zone of prominences mapped during this project is reported to consist of oolitic and pelletal limestone, capped with coral debris (Macintyre and Milliman, 1968 and 1970) and toward the southern end, living coral colonies.

Convolutions seen along the seaward edge of the mapped ridge systems (Figure 2, B and C), as well as visual observations from the J-S-L II, suggest that this outer ridge system may have formed on the framework of an early Holocene reef system. Present day depths at this location (~80 m) correspond to those described by other authors for submerged reefs in the Caribbean, which may indicate a pause in the post-Pleistocene rise of sea level (Macintyre, 1967; Macintyre and Milliman, 1970; Lighty et al., 1978). Depth variation along this ridge, roughly 25 m top to bottom, corresponds to existing depths of reef crest, fore-reef slope, and reef bottom seen today in the Florida Keys.

Topographic features seen along the shoreward side of the study area (Figure 2; A, B and C) typified by the Mounds observation site, appear to be oolitic dune-type formations. The relict faunal composition of these areas reported by Macintyre and Milliman (1970) indicates a shallow water origin. Specifically mentioned are substantial numbers of ghost shrimp (Callianassa sp.) burrows taken in dredge samples. This species is known to occur most commonly in intertidal areas, although it has been reported to be present to depths as great as 10 m (Macintyre and Milliman, 1970).

Macintyre and Milliman (1970) date oolitic limestone samples collected from the shelf edge ridge system between Cape Kennedy and Palm Beach at~ 9600 to~ 13,800 years BP. These dates correspond to others recorded from drowned late Pleistocene or early Holocene dune-ridge systems along Southeastern Florida and in the Caribbean. Lighty et al. (1978) reports that conditions favorable to reef development existed in the western Atlantic during the early Holocene.

Macintyre and Milliman (1968) stated that all sections of the shelf edge ridge complex they examined were inactive and their eroded contours were believed to result from former lower stands of sea level during the Pleistocene. Uchupi (1968) reports many similar ridges and bars from other segments of the Eastern United States continental shelf. He suggests that although these ridges and bars were apparently formed during lower stands of sea level, some of them were probably still active, at least during intense winter storms. Observations made on submersible dives during this study suggest that erosional shaping activity is definitely taking place at this time in the Sebastian Pinnacle System.

Depressions such as Heggerman's Hole are reported from several areas along the Florida continental margin. They may have been formed by the erosive activity of the Florida Current as Pratt

(1966) suggested. However, the actual structure of the sides and floor of the Heggerman's Hole depression suggests a sinkhole-like origin. This feature could have been produced by limestone sapping which occurs in submarine freshwater springs (Manheim, 1967).

## CONCLUSIONS

Despite the fact that the major structural modifications to the Sebastian Pinnacle System probably occurred during the early Holocene, the system remains active today. Sediment transport and erosion by the Florida Current appear to be proceeding at an appreciable rate throughout the whole system. Side-scan sonographs, as well as depth recorder traces, confirm the complexity of this system. Detailed observations using the J-S-L II suggest a coral reef framework for the outer ridge system and an oolitic dune-type origin for the inner or shoreward mound system.

## LITERATURE CITED

Avent, R. M., M. E. King and R. H. Gore, 1977, Topographic and faunal studies of shelf-edge prominences off the central eastern Florida Coast. Int. Revue ges. Hydrobiol. 62 (2): p. 185-208.

Flemming, B. W., 1976, Side-scan sonar: A practical guide. Int. Hydrograph. Rev. L111 (1): p. 65-92.

Lighty, R. G., I. G. Macintyre and R. Stuckenrath, 1978, Submerged early Holocene barrier reef south-east Florida shelf. Nature 276: p. 59-60.

Macintyre, I. G. and J. A. Milliman, 1968, Ridge-like features on the outer shelf and upper slope, continental margin, southeastern United States. (Abs.) Program, Geol. Soc. America Annual Mtg., Mexico City, Mexico, Nov. 11-13, 1968: p. 184-185.

_____, 1970, Physiographic features on the outer-shelf and upper continental slope, Atlantic continental margin, southeastern United States. Bull. Amer. Geol. Soc. 81: p. 2577-2598.

Manheim, F. T., 1967, Evidence for submarine discharge of water on the Atlantic continental slope of the southern United States, and suggestions for further search. Trans. N. Y. Acad. Sci. 29: p. 839-853.

Pratt, R. M., 1966, The Gulf Stream as a graded river. Limnol. Oceanogr. 11: p. 60-67.

Thompson, M. J., L. E. Gilliland and J. E. Mendlein, 1978, Bathymetric mapping of three selected areas on the southeastern Florida continental shelf. Harbor Branch Foundation, Inc. Technical Report No. 027, 54 p.

Uchupi, E. , 1966, Shallow structure of the Straits of Florida. Science 153: p. 529-531.

_____, 1968, Atlantic continental shelf and slope of the United States: Physiography U. S. Geol. Survey Prof. Paper 529C: CL-C38.

_____, 1969, Morphology of the continental margin southeastern Florida. Southeastern Geol. 11: p. 129-134.

# FURTHER OBSERVATIONS ON GLACIATION IN THE SHINING

# ROCK AREA, NORTH CAROLINA: DISCUSSION AND REPLY

## Discussion

### By

Loren A. Raymond
Department of Geology
Appalachian State University
Boone, North Carolina 28608
and
Donald H. Cadwell[*]
New York State Geological Survey
Albany, New York 12230

Pleistocene alpine glaciation of the southern Appalachian Mountains has been regarded as unlikely since Wentworth (1928) briefly considered and rejected the hypothesis. However, Berkland and Raymond (1973) resurrected the alpine glaciation hypothesis following recognition of a U-shaped valley and cirque-like form on Grandfather Mountain in northwestern North Carolina. Soon afterward, Haselton (1973) reported similar features in the Shining Rock and Sam Knob 7.5-minute quadrangles in southwestern North Carolina.

Haselton (1975; 1976; 1979a, 1979b) subsequently reversed his earlier opinion and reported no evidence for alpine glaciation in the Shining Rock quadrangle. We believe Haselton's (1979b) rejection of the glacial hypothesis is premature and ignores his own admonition quoted here from the conclusion of his earlier (1973) paper. "To veterans of alpine glacial geology and geomorphology many of the features discussed above may seem too small and underdeveloped to suggest incipient glaciation. The reader must keep in mind the geographic setting of the southern Appalachians. This region was not covered by an ice sheet as was its counterpart to the north and furthermore, the climate ameliorated here much earlier than it did farther north. Postglacial conditions were warmer, more humid, and wetter so that mass-wasting, weathering and stream action have masked many of the earlier formed features discussed above."

The thrust of Haselton's 1979 paper and his earlier ones (1975; 1976; 1979a) is that none of the features found in well glaciated areas has been recognized in the Shining Rock quadrangle. * Considering Haselton's (1973) comment quoted above and the fact that, with one exception, Berkland and Raymond (1973), Haselton (1973), and Raymond (1975, 1977) only suggested that small cirque glaciers and/or perennial snow fields may have existed at the highest elevations in the southern

Appalachian Mountains, it is not surprising that well developed glacial features cannot be found. Weathering and erosion would have modified the limited features produced during Wisconsinan glaciation. Further, glacial ice may have developed to a maximum at some locations during the Illinoian glacial stage rather than the Wisconsinan stage repeatedly referred to by Haselton (1979b). Thus, 70,000-100,000 years (Goldthwait, 1976) of weathering and erosion could have served to further obscure glacial features not enhanced during the Wisconsinan.

Haselton (1979b) has listed several features found in well glaciated terrains that are not found in the Shining Rock quadrangle. These include U-shaped valleys with step-like profiles and rock basins; moraines; glacial grooves, polish, and striations; whalebacks and roche mountonnees; truncated spurs; ice marginal meltwater channels; eskers, kames, or kame terraces; horns; or tills. Most of these features would not be expected in an area like southwestern North Carolina where incipient glaciation was followed by millenia of weathering and erosion.

Specifically, Haselton (1979b, p. 120) notes that only V-shaped and "pan-shaped" (rather than U-shaped) valleys are found in the Shining Rock area. No profiles are provided, but we presume Haselton refers to valleys with cross sections like those of Yellowstone Prong (Figure 1A) or the head of Greasy (Grassy) Cove Prong (Figure 1B) as "pan-shaped." How were these "pan-shaped" valleys formed? Haselton offers no suggestion, but we suggest that they might be glacially eroded.

With regard to V-shaped valleys, post-glacial erosion may have modified earlier formed profiles. Today, downstream from receding modern glaciers, erosion has modified the U-shaped profile of glacially eroded valleys.

Haselton (1979b, p. 120, 2.) denies that step-like profiles are present in the Shining Rock area. No profiles are presented. Yet, the longitudinal profiles of the valleys of Yellowstone Prong, Grassy Cove Prong, and South Prong of Shining Creek all show stepped profiles (Figure 2). Although we would not argue that all of these valleys necessarily harbored glaciers, Haselton's observation is not valid.

An absence of moraines is noted by Haselton (1979b, p. 120, 5.). However, we would not expect moraines here for three reasons. First temperate ice often does not produce significant moraines, as was the case in many areas of Pennsylvania and New York (Denny and Lyford, 1963). Second, if glaciers had been present in the Shining Rock area they would be small cirque glaciers that would produce no significant

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*Note: Haselton's (1979b) Figure 1 actually covers all of the Shining Rock 7.5-minute quadrangle, not only the west half as indicated in the figure caption. In addition, the longitude shown for the southwest corner of Figure 1 actually represents the longitude of the east edge of the map.
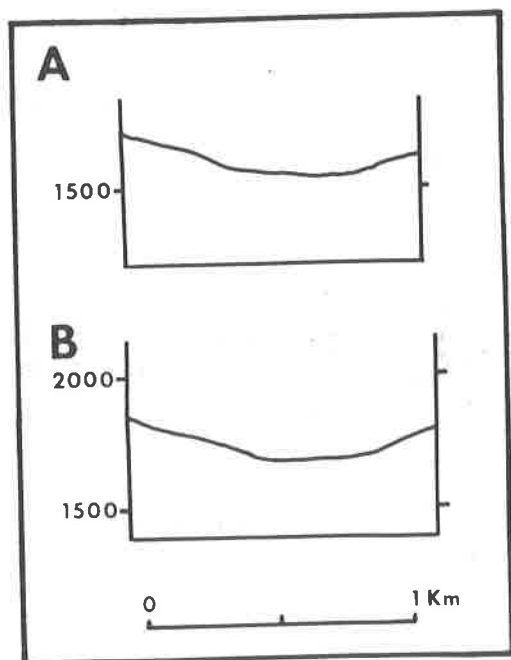
Figure 1. Topographic profiles across valleys in the Shining Rock quadrangle. A - Yellowstone Prong Profile just east of confluence of two main valley head tributaries. B - Profile across valley head of Greasy (Grassy) Cove Prong. Natural profiles. Elevations in meters.

glacial deposition (Boulton, 1974). Third, post-glacial erosion, periglacial activity, and mass wasting would have destroyed any small moraines produced.

Haselton (1979b, p. 120, 6. part 7.) did not find striated, polished, or faceted cobbles or boulders in the area. Although faceted boulders might possibly be preserved, polish and striations would not be preserved in the gneiss, schist, and migmatite bedrock (Haselton, 1973) found in the Shining Rock area. Even facets would probably be destroyed considering expected weathering rates for an area such as this (Owens and Watson, 1979).

Saprolite and colluvium up to 1.8 meters thick were noted by Haselton on valley sides and ridge crests (1979b, p. 120, 8). Figure 3 shows a trail on the highest peak which cuts through soil to bedrock. Haselton suggests that these observations do not support the thesis of alpine glaciation. We feel that the observations do support periglacial or glacial activity.
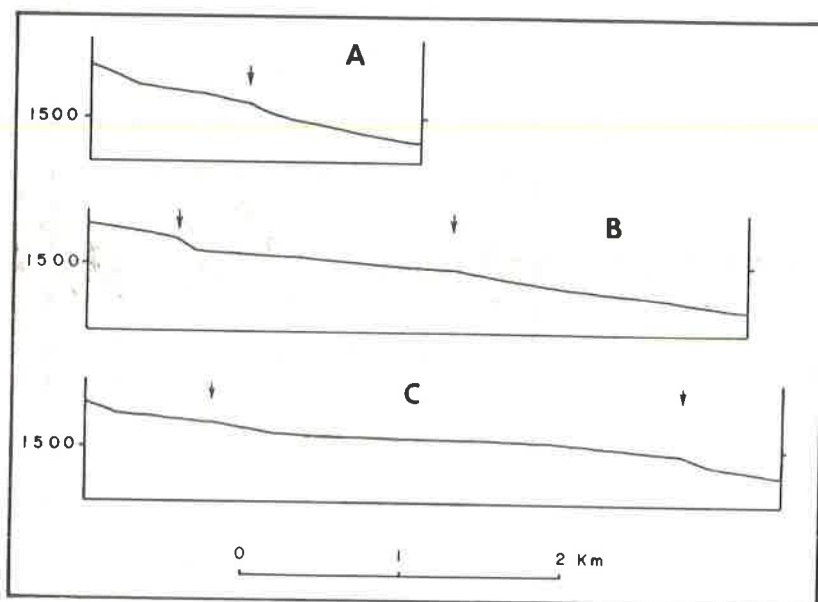
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Figure 2.  Longitudinal profiles along stream channels in the
Shining Rock quadrangle showing stepped nature
of profiles.  A - Yellowstone Prong profile.  B -
Greasy (Grassy) Cove Prong profile.  C - South
Prong of Shining Creek profile.  Arrows indicate
steps.  Profiles are natural with elevations in
meters.

A soil of 1.8 meters is thin compared to soils developed in non-
glaciated areas of the Grandfather Mountain area.  Near Boone, North
Carolina, soils on non-glaciated ridge crests range up to 15 meters
thick.  Considering the rates of soil formation in an area of similar
temperatures and rainfall to the Shining Rock area described by Owens
and Watson (1979), it is possible that the thin soils observed by Hasel-
ton on ridge tops formed in less than 100,000 years.  Certainly, the
marked disparity between soil thicknesses on ridge crests at higher and
lower elevations in the southern Appalachians needs further study.

Point 9 raised by Haselton (1979b), regarding the absence of
ice-erosional or depositional forms at an elevation of 915m (3000 feet)
seems to be a "straw man." Only Harrington (in Haselton, 1979b) has
proposed that glaciation extended below 1067 meters (3500 feet), and
that proposal has not been documented in published form.

The stoss and lee topography and roche moutonnees, absent in
the Shining Rock area (Haselton, 1979b, p. 122, 10) are more common-
ly found associated with continental glaciers.  They would probably not
be formed by the small cirque glaciers envisioned by Berkland and

Raymond (1973) or Raymond (1975, 1977). Whaleback forms might be formed and preserved (Raymond, 1977).

Like roche moutonnees and U-shaped valleys, truncated spurs would not be produced by the small glaciers that perhaps existed in the Shining Rock area. The glaciers would be too small to generate major erosion and no major spurs (to be truncated) are evident in the cirque-like forms we observed on the Shining Rock quadrangle.

Considering the size of the proposed glaciers, massive amounts of detrital materials would not be expected in major valleys or along valley walls. Haselton (1979b) suggests that they would be. However, we concur with Haselton (1979b) that whatever materials were eroded would produce valley fills--perhaps like those we note at John Rock, Yellowstone Prong, near the Pink Beds, and elsewhere in the quadrangle. That these deposits would be trenched and terraced is moot, especially considering the fact that the valley fills typically formed behind bedrock constrictions downstream. We believe that valley fills and thin upland soils may support rather than contravene the glacial hypothesis.

The absence of ice-marginal bedrock channels, like the absence of other small scale features, is not surprising considering post-glacial weathering and periglacial activity. If such channels had formed, they would be small features easily destroyed or masked by later erosion--especially in less resistant rock types like the schists found in the Shining Rock quadrangle.

Haselton (1979b, p. 126, 14.) also notes that no till has been observed in the Shining Rock area. We concur that till should be present, but to conclude that none eixsts prior to completion of detailed mapping and fabric studies and electron microscope analysis of sand grain textures in diamictons is premature.

Haselton's (1979b) final paragraph before his conclusion is a collection of non-sequiturs, indefensible assertions, and statements of the obvious. Sentence one asserts that "snowbanks or snowfields seldom persisted throughout the summer season." No data are presented here or elsewhere in the paper to support such a conclusion. Sentence three suggests that if glaciers had been present, horns would have formed. Even in the high Rocky Mountains of Colorado, where glaciers are still present, horns are not ubiquitous, and they certainly would not be expected in the southeast where glaciation was incipient. Haselton (1979b) states this obvious point in sentence four. Why, then, has he raised the question if not simply to create another "straw man" to knock down? We do not concur that an ice cap would produce a landscape that, either initially or after 20,000-100,000 years of weathering and erosion, would necessarily be "covered with glacial cobbles and boulders" (Haselton, p. 126). To our knowledge, no one, including Haselton (1973), has suggested ice caps for the region. Finally, Haselton presents a photograph (his Figure 8) of a valley head that clearly lacks a well developed cirque-like form. This is the most convincing

piece of evidence presented in the paper. However, the valley selected is a south facing valley head on a ridge extending east from Black Balsam Knob--not an east facing valley head as one might assume from reading the ambiguous figure caption. The valley head selected is one of the less well developed cirque-like forms in the Black Balsam Knob area and does not have topography that precludes the existence of pre-Wisconsinan glaciation or a small perennial Wisconsinan snow or ice field at its head.

Haselton's paper (1979b) essentially demonstrates that none of the features found in extensively glaciated areas such as the Colorado Rockies or Swiss Alps are found in the Shining Rock quadrangle. This conclusion is not surprising, as none of the published literature suggesting Pleistocene glaciation of the Southern Appalachian Mountains suggests more than a single small valley glacier, a few cirque glaciers, and perennial snow fields-- an incipient form of glaciation (Berkland and Raymond, 1973; Haselton, 1973; Raymond, 1977). What is surprising is that Haselton (1979b) has developed such a cursory, poorly documented case against incipient glaciation, when what is needed are detailed geomorphic and sedimentological analyses. The possibility of glacial modification of the Shining Rock landscape has not been disproved.

## REFERENCES CITED

Berkland, J. O., and Raymond, L. A., 1973, Pleistocene glaciation in the Blue Ridge Province, southern Appalachian Mountains, North Carolina: Science, v. 181, p. 651-653.

Boulton, G. S., 1974, Processes and pattern of glacial erosion: in D. R. Coates, ed., Glacial Geomorphology: Publications in Geomorphology, State Univ. of New York, Binghamton, p. 41-88.

Denny, C. S., and Lyford, W. H., 1963, Surficial geology and soils of the Elmira-Williamsport region, New York and Pennsylvania: U. S. Geol. Survey Prof. Paper 379, 60 p.

Goldthwait, R. P., 1976, Paleosols and Quaternary History of western Ohio: in W. C. Mahaney, ed., Quaternary stratigraphy of North America: Dowden, Hutchinson, and Ross, Inc., Stroudsburg, Pennsylvania, p. 159-160.

Haselton, G. M., 1973, Possible relict glacial features in the Black Balsam Knob and Richland Balsam area, North Carolina: Southeastern Geology, v. 15, p. 119-125.

_____, 1975, Observations regarding Wisconsin glaciation in the Shining Rock quadrangle, North Carolina (abs.): in The Second Annual Conference on the Quaternary History of the Southeastern United States (Abstract Volume): Appalachian State University, Department of Geology, Boone, N. C., p. 10.

Haselton, G. M., 1976, Wisconsin glaciation in the southern Appalachians, North Carolina: Geol. Soc. of America Abs. with Programs, v. 8, p. 192.

_____, 1979a, Further observations on glaciation in the Black Balsam Knob and Richland Balsam area, North Carolina: in L. A. Raymond, ed., Proceedings of the Second Annual Conference on the Quaternary History of the Southeastern United States: Appalachian State University, Department of Geology Publication No. 1, p. 7-17 (in press).

_____, 1979b, Further observations on glaciation in the Shining Rock area, North Carolina: Southeastern Geology, v. 20, p. 119-127.

Owens, L. B., and Watson, J. P., 1979, Landscape reduction by weathering in small Rhodesian watersheds: Geology, v. 7, p. 281-284.

Raymond, L. A., 1975, Were Pleistocene glaciers present in the southeastern United States? (abs.): in The Second Annual Conference on the Quaternary History of the Southeastern United States (Abstract Volume): Appalachian State University, Department of Geology, Boone, N. C., p. 11.

_____, 1977, Glacial, periglacial, and pseudo-glacial features in the Grandfather Mountain area, North Carolina: Southeastern Geology, v. 18, p. 213-229.

Wentworth, C. K., 1928, Striated cobbles in southern states: Geol. Soc. of America Bull., v. 39, p. 941-954.

Reply

By

George M. Haselton
Department of Geology
Clemson University
Clemson, South Carolina 29631

It was rewarding to find a critical discussion of my article in Southeastern Geology, Haselton (Vol. 20, No. 2, March, 1979).

Many aspects of late Pleistocene and Holocene geomorphology have been neglected in this part of eastern United States and Berkland and Raymond's initial article about Grandfather Mountain Glaciation, (1973) sparked much renewed interest and considerable controversy.

My recent field studies in the upper portions of the Davidson and East Fork of the Pigeon River drainages were simply to point out what I have not been able to demonstrate in the field to date; namely positive evidence for alpine glaciation. I would be delighted if any investigators could adequately document a case for glaciation. Very

understandably one cannot expect the extreme Swiss Alpine or Rocky Mountain glacial forms in a topographic and geographic setting such as the one under consideration, and I am sure my colleagues at Appalachian State are fully aware of this.

One would certainly not expect to find evidence of Illinoian Glaciation in the region under consideration since the time span for weathering and erosion has been too great. I have directed my comments to the Wisconsinian Glacial Age. The bounding margins of the Wisconsin continental glaciers were only 50 - 100 miles (80 - 160 km) short of those of the Illinoian. Furthermore, this advance took place much more recently. The climatic change enhanced by Wisconsin continental glaciation was most certainly felt this far south.

Indeed, as everyone knows, we do find residual masses of till from earlier glaciations within the known glacial limit north of us. If till exists in the study area under consideration, it should be exposed somewhere along the numerous highway cuts of the Blue Ridge Parkway or in deep stream channels. I would be most happy to see a "bona fide" till in this region.

Profiles across and parallel to valleys can be very misleading. One can "make them" do a number of things. Even longitudinal profiles made within ten miles of our Clemson Campus by students and professors can show the same frequent nickpoints or thresholds pointed out by Professors Cadwell and Raymond. Cross valley strike of resistant rock units can provide numerous "pseudo-glacial thresholds". What the trained field geologist is looking for are significant or substantial erosional closures in the upper portions of these high altitude valleys. I have, as yet, not found them and hence did not attempt to add cross-sections to my paper.

Mass-wasting alone can produce a number of "pan-shaped" cross valley profiles; it is not necessary to call on glaciation.

I disagree with Cadwell and Raymond; if a well developed U-shaped valley exists, and one can prove it was developed by glacial erosion, even post-Wisconsin weathering and mass-wasting should not have filled the entire valley with weathered detritus to its upper reaches.

No indeed, today, downstream from modern glaciers one can still see the upper portions of older U-shaped valley profiles rather clearly.

Temperate ice can produce moraines as has been demonstrated recently in the Adirondacks, Green Mountains, and portions of the Appalachians in Maine. This information is based on recent discussions with colleagues in the field.

Fine striations produced by faulting can be found on quartz-rich facies of gneissic outcrops in selected areas of the Shining Rock region. As a matter of fact, a considerable portion of this region is underlain by granite-gneiss which is a resistant rock unit "holding up" most of the high peaks. Any glacial striae developed on this rock type, and

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protected by till, should still be visible today. Unfortunately no glacial striations have as yet been reported by my students, staff or other associates. A point of interest that should be brought to the reader's attention is the fact that very delicate slickensides are still well preserved here in our piedmont saprolites. This being the case, one should expect to find striations on the resistant gneissic outcrops in the high Blue Ridge if they were made by post-Illinoian glacial ice.

Soils can thicken abruptly on the interfluves between steep-walled stream valleys, and if a study was executed simply to find soil depths greater than 1.8 meters, I doubt that this would be a very formidable task.

Point 9, raised in my paper, simply is asking the reader if we know if snowline during the last glacial maximum extended to an elevation as low as 3,000 feet (915 m).

Good stoss and lee features can easily be produced in valley bottoms by local alpine glaciers. Since these kinds of features were not seen one has to assume there was no ice there to produce them.

I am pleased that Professors Cadwell and Raymond do agree that trenched valley fills would be supportive evidence for former glaciation. The fills (alluvium) now seen at John Rock and the upper portion of Yellowstone Prong appear to me to be local Holocene sediments.

With careful study of the uppermost contours, immediately south and north of Black Balsam Knob one can see that the upper rim of these ravines face eastward. Oddly enough the best "cirque-like" forms face south.

Potholes in any glacial marginal channels can exist for thousands of years after deglaciation, especially in resistant rock such as granite-gneiss.

To create the inference of an "assertion", Cadwell and Raymond misquoted a statement of mine. In my final paragraph before conclusions I stated, "It would seem that snowbanks or snowfields seldom persisted throughout the summer season." Those four critical words "It would seem that" make a considerable difference in the interpretation (Haselton, 1979, p. 126).

Professor Raymond was a reader of the original draft of this paper, as cited under acknowledgements on page 120. At the time of my first writing, he took the time and effort to make a number of suggestions. I am quite sure that coming back for a "second helping" is to point out to the reading audience the magnitude and importance of this topic that both he and I have been laboring with for a number of years. I do appreciate his support and continued deep interest and that of his colleague Professor Cadwell. With continued positive cooperation, I am very hopeful that we can someday resolve this Southern Appalachian Glaciation question.

My conclusions, based on my field observations, remain unchanged.