ORIGIN OF ORE DEPOSITS IN THE CARTERSVILLE MINING DISTRICT & STRATIGRAPHIC AND KINEMATIC EVIDENCE FOR THE SEPARATION OF THE CARTERSVILLE-GREAT SMOKY AND EMERSON-TALLADEGA FAULTS

EDITED BY:
RANDY L. KATH AND KAREN S. TEFEND

49TH ANNUAL FIELD TRIP
OF THE GEORGIA GEOLOGICAL SOCIETY
CARTERSVILLE, GEORGIA, OCTOBER 9-11, 2015
OFFICERS OF THE GEORGIA GEOLOGICAL SOCIETY

2015

President ----------------------- Timothy M. Chowns
Department of Geosciences
University of West Georgia
Carrollton, Georgia 30118

President Elect----------------- Open

Past President ----------------- Julian Gray
Tellus Science Museum
Cartersville, Georgia 30120

Secretary ---------------------- Randy L. Kath
Department of Geosciences
University of West Georgia
Carrollton, Georgia 30118

Treasurer ---------------------- Karen S. Tefend
Department of Geosciences
University of West Georgia
Carrollton, Georgia 30118

2015 FIELD TRIP

Leaders: Jim DeCinque, Vulcan Materials Company
         Rodney Matiak, Vulcan Materials Company
         Stanley Bearden, New Riverside Ochre
         Timothy Chowns, University of West Georgia
         Thomas Crawford, University of West Georgia
         Randy Kath, University of West Georgia

The Georgia Geological Society is a non-profit organization
Incorporated in the State of Georgia

Georgia Geological Society Guidebooks are published by:
The Georgia Geological Society, Inc., Carrollton, Georgia
ORIGIN OF ORE DEPOSITS IN THE CARTERSVILLE MINING DISTRICT & STRATIGRAPHIC AND KINEMATIC EVIDENCE FOR THE SEPARATION OF THE CARTERSVILLE-GREAT SMOKY AND EMERSON-TALLADEGA FAULTS

EDITED BY:
RANDY L. KATH AND KAREN S. TEFEND

49TH ANNUAL FIELD TRIP OF THE GEORGIA GEOLOGICAL SOCIETY CARTERSVILLE, GEORGIA, OCTOBER 9-11, 2015

GEORGIA GEOLOGICAL SOCIETY GUIDEBOOKS VOLUME 34, NUMBER 1 OCTOBER 2015
CONTENTS

VULCAN MATERIALS BARTOW QUARRY: BARTOW COUNTY, GEORGIA
by: Jim DeCinque and Ken Nelson ................................................................. 1

MINERALOGY, GEOTHERMOMETRY, AND GEOBAROMETRY OF THE ROWLAND SPRING FORMATION, ALLATOONA COMPLEX, GEORGIA
by: Randy Kath and Thomas Crawford ............................................................ 7

A REVIEW OF ORE DEPOSITS IN THE CARTERSVILLE DISTRICT, BARTOW COUNTY, GEORGIA, INCLUDING RELATED DEPOSITS IN POLK AND FLOYD COUNTIES
by: Timothy Chowns and Randy Kath ............................................................. 25

DAY 1 ITINERARY- ALLATOONA COMPLEX AND OCOEE SERIES ROCKS ........................................ 35
STOP 1. ROWLAND SPRING FORMATION TYPE LOCALITY. RANDY KATH AND TOM CRAWFORD
STOP 2. CORBIN METAGRANITE. JIM DECINQUE, RODNEY MATIAK, JESSICA SCOTT, AND BRANDON KING
STOP 3. OCOEE METASEDIMENTS AND YELLOW BREECHES MEMBER (?) OF THE WILHITE FORMATION; ALLATOONA DAM ABUTMENT. RANDY KATH, TOM CRAWFORD, AND BILL WITHERSPOON
STOP 4. GEOLOGY OF THE IRON HILL CAMPGROUND AND WALKING TRAIL AT ALLATOONA LAKE, BARTOW COUNTY, GEORGIA: STRATIGRAPHIC AND KINEMATIC EVIDENCE FOR SEPARATION OF THE CARTERSVILLE-GREAT AND EMERSON-TALLADEGA FAULTS. RANDY KATH AND TOM CRAWFORD

DAY 2 ITINERARY- CARTERSVILLE MINING DISTRICT ............................................................................ 49
STOP 1. NEW RIVERSIDE OCHRE’S EMERSON, GEORGIA, BARITE MINE. STAN BEARDEN
STOP 2. COOPER FURNACE AND OCOEE SUPergroup. TIM CHOWNS AND BILL WITHERSPOON

GEORGIA GEOLOGICAL SOCIETY PUBLICATIONS ................................................................................. 57

PLATE 1- GEOLOGIC MAP OF THE CARTERSVILLE MINING DISTRICT, GEORGIA- INCLUDING THE CARTERSVILLE, ALLATOONA DAM, BURNT HICKORY RIDGE AND ACWORTH (NORTHERN) 7.5-MINUTE QUADRANGLES, GEORGIA
by Randy Kath, Thomas Crawford, John Costello, and Stanley Bearden ..................... 63

Enjoy the trip!
DEDICATION

The Georgia Geological Society would like to dedicate this year’s annual fieldtrip and guidebook to John O. Costello.

John was born in Atlanta, Georgia on January 20, 1947. He was the oldest son of the late Walter O. Costello, of Macon, GA, and the late Joann W. Costello, of Atlanta, GA. John is survived by his former wife and mother of his three children, Alice J. Costello of Atlanta, Georgia, and their children Oliver W. Costello of Gainesville, Florida; Katherine C. Braxton of Atlanta, Georgia; and Molly C. Sanford of Memphis, Tennessee; along with four grandchildren Paige, Grady, and twins Mae and Harper; and surviving siblings Michael H. Costello of Atlanta, Georgia; Christine M. Costello of Atlanta, Georgia; Annmarie C. Aquino of New York City, New York; and Clare M. Costello of Atlanta, Georgia.

John graduated in 1965 from Sandy Springs High School. John was a graduate of Georgia State University, where he received a B.S. in geology with an art minor, and the University of South Carolina, where he received his masters and Ph.D. degrees. John began his professional career with the Georgia Geological Survey (GGS) in 1973 and helped compile the 1976 State Geologic Map and mapped geology in the Georgia Blue Ridge, Piedmont, and Valley and Ridge Provinces. John has also worked for the US EPA, several environmental consulting companies, and The Georgia Marble Company where he was Mineral Resources Director for six years. John was an Adjunct Professor of Geology at Georgia State University and the University of West Georgia where he taught a field-oriented class on Georgia Geology. Since 2002, John had been with the GGS where he conducted geologic mapping and supervised statewide mining activities. John has authored over 35 abstracts and papers on southern Appalachian geology. He was an active member and former President of the Atlanta Geological Society and Georgia Geological Society. He retired from the State of Georgia in 2013.

John had many interests away from geology. Aside from his love of the natural sciences, John loved music and was the drummer for numerous bands as the Sandy Spring Beatles, The Majestics, and Rude Cooty. His art interests encompassed drawing, painting, and photography, but for the last 35 years, his primary medium had been stained glass. He was also Boy Scout Troop Leader to BSA Troop 87 of Roswell, Georgia for six years. He enjoyed the great outdoors where he camped, hiked, and fished.
ACKNOWLEDGMENTS

As always the Society wishes to acknowledge a number of companies and individuals for their support and assistance in preparation for this year’s fieldtrip. First, of course, we thank Randy Kath and Tim Chowns for agreeing to lead and organize the trip and to edit the guidebook. They have been ably supported by the other fieldtrip leaders, especially Stanley Bearden and Tom Crawford. The guidebook was printed at the University of West Georgia. Fieldtrip registration and logistics were handled by Karen Tefend and Tim Chowns. Additionally, we would like to thank Matthew Howe and Ginny Mauldin-Kenney for helping with fieldtrip logistics.

Without the support and corporation of New Riverside Ochre, Stanley Bearden; U.S. Army Corp of Engineers, Rusty Simmons; Red Top Mountain State Park, Mr. Kelly Howington; and Vulcan Materials Inc, Mr. Jim DeCinque and Rodney Matiak we would not have been able to conduct the detailed geologic mapping necessary for this guidebook. Additionally, we would like to thank all of the other land owners for allowing access to properties.

The Friday night social is an important part of the fieldtrip weekend. This year financial support was generously provided by friends of John O. Costello for the social. Additionally, the guidebook and fieldtrip are dedicated to John.

We hope everyone enjoys the trip. Several generations of geologists have contributed to the conclusions presented in this guidebook. Many of them are attending this meeting to discuss various interpretations. We can all look forward to a stimulating and fun weekend.

INTRODUCTION

Once again the Georgia Geological Society returns to one of the most complex and enigmatic regions of Georgia geology- the eastern boundary and southern boundary of the Paleozoic sedimentary rocks of northwest Georgia; where they are in a faulted relationship with igneous and metamorphic rocks of the Piedmont/Blue Ridge.


Even a casual review of the geologic maps and commentary provided as guides for these field trips shows emphatically that not all of us can be right all of the time. Unfortunately (?) none of the “discussions” on the outcrop were recorded for posterity. These were often intense; always entertaining, especially to the non-participants; and perhaps sometimes, even productive.

Published literature on various aspects of this area is voluminous; and a list of “disagreements” would be longer than a list of “agreements”. Considering this history, perhaps you will be pleased to note that, presented here, for the 2015 Georgia Geological Society Fieldtrip, is the most recent “final word” on the geology and origin of ore deposits of this profoundly intriguing enigma.

From several who have been there- and had to change an interpretation a time or two. Until next time . . .
HISTORY

As early as 1962, the Georgia Highway Department, known now as the Georgia Department of Transportation, or DOT, had developed an interest in locating sources of siliceous rock in or close to the Paleozoic area of northwest Georgia. The rock would not only be required to meet all specifications for aggregate that limestone must meet, but it would have to be comprised of silica in proportions sufficient to produce skid resistant values adequate to permit its use in road surfaces.

Research by both Federal and State Transportation systems had indicated that limestone aggregate, when not used in proper combinations with siliceous aggregate, would cause slippery roads and was a contributing factor in many traffic accidents. With the onset of the construction of the interstate highway system which would bring with it a higher traffic volume at increased speeds, it was mandatory that every possible safety measure be built into the vast network of new roads, and slippery pavement surfaces could not be tolerated. It was not until 1972, however, that the DOT, following Federal guidelines, required siliceous material to be used as the fine aggregate in Portland cement concrete surfaces while permitting the use of limestone as the coarse aggregate. Just the opposite would apply for asphaltic concrete surfaces. The coarse aggregate must be comprised of skid resistant material while limestone would be permitted for use as fine aggregate.

In the mid 1960’s the DOT’s geology section began an extensive exploration program for skid resistant materials in NW Georgia, and at the conclusion, most sandstone deposits in this part of the state had been sampled and tested. Unfortunately, rock in all but a few of the deposits was found to be too friable and failed to meet the L.A. degradation specifications for coarse aggregate, and were to fine too meet the gradation specifications for fine aggregate. For the few deposits that passed the L.A. degradation specifications, there were other shortcomings such as poor location, scant reserves, and soundness failures.

Consequently, the search zone was narrowed to the rocks SE of “the fault” especially in the Bartow-Cherokee Counties area where extensive outcrops (mainly in the form of boulders) of Corbin metagranite (Higgins and others, 1996) occur. In 1973, Vulcan Materials began looking at rocks in this formation to supply a skid resistant stone for use in the construction of the outer loop which was already in the rumor stages and other potential projects in northwest Georgia.

By 1976, a site had been located, drilled, and leased, but seven years and two division presidents later, the lease was allowed to expire without exercising an option to renew. Five
years later in 1988, the Vulcan Geology Department was again directed to locate a prospect in the same general area as the tract that was leased in 1976, and was given very limited time to determine the best potential quarry within a 1200 acre area.

Exploration

The Bartow County quarry site was optioned from the Inland-Rome Paper Company during the summer of 1988. Vulcan was granted a 90 day option to explore 1200 acres located on the north and south sides of Wilderness Camp road (see Figure 1). Exploration of the site was difficult due to the short option period, large size of the property, limited access, and rugged terrain. Early reconnaissance of the site indicated that core drilling would be the most effective tool for exploration at the site.

The following problems had to be addressed within the limited time period.

- What are the physical and chemical properties of the Corbin metagranite and the rocks associated with the deposit?
- Where is the “best” location for the initial pit development and future expansion area?
- What is the thickness of overburden and weathered rock throughout the defined pit area?

Vulcan moved three core drills onto the site in late July 1988. Two of the three core drills were operated around the clock, seven days a week. A total of 72 core holes were placed over the property. The “B” sized cores were logged on site, bagged, and transported to the Vulcan Materials Research and Development Laboratory in Birmingham, Alabama. Physical testing performed on the core included the Los Angeles degradation test, specific gravity, absorption, magnesium sulfate soundness loss, and petrographic analysis. Chemical tests were completed by wavelength dispersive X-ray analysis (XRF). A description of these tests and results from the testing are included in Table 1 and Figure 2.

During the fall of 1988 a competitor company announced plans to open a new quarry approximately 2.5 miles east of Vulcan’s site. Public interest and debate quickly escalated around the competitors new site. Vulcan decided to cease all drilling and exploration activities until issues affecting the competitors site could be addressed.

As the Legal Department pursued zoning and environmental issues, the Geology Department continued to process the data generated from the short drilling period. Results from the drilling indicated that an area located in the northeast section of the property contained the least amount of overburden and weathered rock. An important distinction is made between overburden and weathered rock in the quarry business. Overburden is generally defined as the loose material that can be removed by pans, bulldozers or loaders, while weathered rock is defined as the consolidated rock that must be blasted to remove. The two are separated since the cost to blast and remove weathered rock is considerably higher than removing overburden.

The northeast area contained a sizable amount of Corbin metagranite. Most of the property to the south contained a large amount of other rock types associated with the porphyritic phase of the Corbin metagranite. Physical test results from the core drilling indicated the porphyritic metagranite would make an excellent construction aggregate. However, chemical tests indicated high iron oxide content was associated with the reddish-brown to orange granofel (Higgins and others, 1996) bodies scattered throughout the site. The high iron content was suspected to be related to the abundant pyrite and hematite visibly seen in the core. The Georgia Department of Transportation requires that all coarse aggregate used in Portland cement contain less than 0.01 percent total sulfide sulfur.
Figure 1. Location of Vulcan Materials Bartow Quarry
During the winter of 1991-92 the Geology Department was requested to begin a new exploration program on the Bartow County property. Access to the northern section of the property from Georgia Highway 20 had been purchased. Feasibility studies showed the cost of constructing an entrance across the property was unusually expensive. To shorten the proposed access road and thus reduce costs, the geologists were asked to concentrate their exploration on the northwest half of the property. Realizing this was not the best location for the pit, an attempt was never the less made to prove an area large enough to quarry. A total of 64 new core holes were placed in this area. The core holes ranged from 50 to 400 feet deep. Detailed geologic mapping (1” = 100’) was completed on the northern part of the property. Core drilling was used to prove the existence of one large fault and several sulfide-rich, granofel bodies located during the construction of the surface geology map. Due to the abundant amount of sulfide-rich

<table>
<thead>
<tr>
<th>Property</th>
<th>Cgn1</th>
<th>Cgn2</th>
<th>Cgn3</th>
<th>Cgn4</th>
<th>Fgn5</th>
<th>Fgn6</th>
</tr>
</thead>
<tbody>
<tr>
<td>Appar. S.G.</td>
<td>2.703</td>
<td>2.721</td>
<td>2.715</td>
<td>2.725</td>
<td>2.862</td>
<td>2.753</td>
</tr>
<tr>
<td>Bulk S.G.</td>
<td>2.673</td>
<td>2.682</td>
<td>2.673</td>
<td>2.680</td>
<td>2.753</td>
<td>2.730</td>
</tr>
<tr>
<td>SSD S.G.</td>
<td>2.684</td>
<td>2.696</td>
<td>2.689</td>
<td>2.697</td>
<td>2.792</td>
<td>2.738</td>
</tr>
<tr>
<td>Absorp.%</td>
<td>0.4</td>
<td>0.5</td>
<td>0.6</td>
<td>0.5</td>
<td>1.4</td>
<td>0.3</td>
</tr>
</tbody>
</table>

**L.A. Degradation**

<table>
<thead>
<tr>
<th>% Loss</th>
<th>Cgn1</th>
<th>Cgn2</th>
<th>Cgn3</th>
<th>Cgn4</th>
<th>Fgn5</th>
<th>Fgn6</th>
</tr>
</thead>
<tbody>
<tr>
<td>27.3</td>
<td>24.1</td>
<td>24.9</td>
<td>23.3</td>
<td>18.4</td>
<td>22.1</td>
<td></td>
</tr>
</tbody>
</table>

**MgSO₄ Soundness**

<table>
<thead>
<tr>
<th>% Loss</th>
<th>Cgn1</th>
<th>Cgn2</th>
<th>Cgn3</th>
<th>Cgn4</th>
<th>Fgn5</th>
<th>Fgn6</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.9</td>
<td>1.7</td>
<td>0.9</td>
<td>0.9</td>
<td>9.4</td>
<td>4.1</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Compound</th>
<th>Cgn1</th>
<th>Cgn2</th>
<th>Cgn3</th>
<th>Cgn4</th>
<th>Fgn5</th>
<th>Fgn6</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>67.64</td>
<td>66.13</td>
<td>66.57</td>
<td>66.52</td>
<td>58.11</td>
<td>64.37</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>15.22</td>
<td>15.61</td>
<td>15.44</td>
<td>14.65</td>
<td>19.29</td>
<td>14.67</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>4.80</td>
<td>4.75</td>
<td>5.08</td>
<td>5.28</td>
<td>12.02</td>
<td>4.38</td>
</tr>
<tr>
<td>CaO</td>
<td>2.06</td>
<td>1.79</td>
<td>1.82</td>
<td>1.06</td>
<td>0.28</td>
<td>2.31</td>
</tr>
<tr>
<td>MgO</td>
<td>1.40</td>
<td>1.77</td>
<td>1.84</td>
<td>2.19</td>
<td>1.22</td>
<td>2.16</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.98</td>
<td>3.38</td>
<td>2.49</td>
<td>4.26</td>
<td>3.14</td>
<td>5.17</td>
</tr>
<tr>
<td>K₂O</td>
<td>4.70</td>
<td>5.18</td>
<td>5.06</td>
<td>4.15</td>
<td>4.32</td>
<td>5.00</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.65</td>
<td>0.74</td>
<td>0.79</td>
<td>0.80</td>
<td>1.17</td>
<td>0.84</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.17</td>
<td>0.20</td>
<td>0.18</td>
<td>0.11</td>
<td>0.08</td>
<td>0.13</td>
</tr>
<tr>
<td>MnO</td>
<td>0.13</td>
<td>0.11</td>
<td>0.11</td>
<td>0.10</td>
<td>0.29</td>
<td>0.10</td>
</tr>
<tr>
<td>Sr</td>
<td>0.04</td>
<td>0.03</td>
<td>0.04</td>
<td>0.02</td>
<td>0.01</td>
<td>0.04</td>
</tr>
<tr>
<td>Zr</td>
<td>0.05</td>
<td>0.05</td>
<td>0.06</td>
<td>0.06</td>
<td>0.07</td>
<td>0.07</td>
</tr>
<tr>
<td>SO₃</td>
<td>0.16</td>
<td>0.26</td>
<td>0.52</td>
<td>0.80</td>
<td>----</td>
<td>0.76</td>
</tr>
</tbody>
</table>

**Sample Descriptions:**

Cgn #1-4  Corbin metagranite
Fgn #5  Sulfide-bearing granofel
Fgn #6  Sheared Corbin metagranite
granofels, drilling was terminated on the northwest property.

Core drilling resumed on the northeast section of the property in late July of 1992. An additional 50 core holes were added to the new section of the property. Special attention was paid to the thickness and orientations of the sulfide-rich granofel bodies in this area. The coarse-grained cores were collected and tested separately from the fine-grained cores.

Quantification and structural relationships of the granofels and shear zones was attempted by using the computer drafting program Autocad to extrapolate between core holes, which were located approximately on 400-foot centers. Very little correlation between the core holes could be drawn. One section of the property was drilled on 100-foot centers in an attempt to better correlate some of the granofel zones. Even on this close spacing, limited correlation could be developed between many of the holes. Core drilling continued at this location until a large sized pit area could be verified. Vulcan Materials requires a minimum of 50 years of reserves at a production rate of one million tons per year.

**QUARRY DEVELOPMENT**

During the Fall of 1994, clearing for the present plant and pit site began. The entrance road was constructed along with the building of the office, scales, shop and plant. The primary crusher is a Hewitt Robins 42 x 48 jaw crusher. The plant production is presently 650 ton per hour for 12 hours per day; however, the plant has the ability to produce 1400 tons per an hour for 12 hours per day. The plant was constructed to specifically allow the production staff to crush and stockpile the “good” clean rock, while if necessary, to crush and stockpile the high sulfide granofels in a separate area on the yard. The equipment for rock removal in the pit includes an Liebherr 6 cubic yard shovel, four International 350 - 50 ton haul trucks and one 155 Komatsu bulldozer. The plant area utilizes one L-180 Michigan loader for customer truck loading and uses one water truck and motor grader for road maintenance.

Most of the aggregate will be used in the production of asphalt and concrete. Special emphasis is placed on the production of stone sizes such as 67, 57, 89, 5, 7, and 810’s. Other products include manufactured sand, surge-stone, Type 1 or 3 rip-rap, and crusher run. The Bartow quarry is now the best located producer and supplier of granitic aggregate in northwest Georgia.

**SUMMARY**

In September of 1994, twenty-one years after Vulcan Materials expressed interest in developing a quarry in Bartow County, the Bartow quarry began operations. Although this was an extraordinarily difficult undertaking for the geology department, in the end, exploration, core drilling, and reserve evaluation may have been one of the easiest parts of the project. During the five years between 1983 and 1988, for example, zoning, air and water permits, and environmental requirements had become stricter and more difficult to obtain. Furthermore, an archeology study had to be conducted to ascertain that no Indian artifacts would be disturbed by a quarrying operation. Additionally, because of the many hills and curves along Georgia SR 20, the location and acquisition of the quarry access road was no small task.

Quarry operators preparing to open a new site face many challenges today such as zoning, environmental regulations, community relations, protection of historic artifacts, wetland issues, construction permits, Federal air and water quality issues, noise and vibration monitoring, and traffic management. Quarry locations are usually picked based on their proximity to the market area. The costs of transportation have a great influence on the location of the chosen site. The quarry site selected by the operator is not always based on the “best” quality of reserves, but is based on the “best” quality rock minable in the market area. At times, many of these issues tend to overshadow the importance of good exploration and quality testing programs.
### SPECIFIC GRAVITY AND ABSORPTION OF FINE AND COARSE AGGREGATE:

The specific gravity test is determined by the ASTM C 128 test method. These tests are based on Archimedes’ principal that a body, when immersed in water is subjected to a vertical upward force equal to the weight of water it displaces. The commonly used specific gravity measurements used in the testing of aggregate are as follows:

- **Apparent specific gravity:** the ratio of the weight of dry aggregate to the weight of water having a volume equal to the solid volume of aggregate excluding its permeable pore space.
- **Bulk specific gravity:** the ratio of weight of dry aggregate to the weight of water having a volume equal to the volume of the aggregate including both its permeable and impermeable pore spaces.
- **Bulk specific gravity - saturated, surface dry (SSD):** the saturated, surface dry specific gravity is the ratio of the weight of aggregate, including the weight of water it contains when permeable voids are saturated, to the weight of an equal volume of water.

The absorption test is determined by the ASTM C 127 test method.

- **Percent water absorption:** is the weight of saturated, surface dry aggregate minus the weight of the same oven dried aggregate divided by the oven dried aggregate weight. Multiply by 100 to obtain the percent absorption value.

### LOS ANGELES DEGRADATION TEST:

The Los Angeles degradation test is determined by the ASTM C 131 test method. The Los Angeles degradation test is the most widely specified test for the evaluation of the resistance of coarse aggregates due to abrasion and impact. A screened and weighed sample is placed in a revolving steel drum with 6 to 12 steel balls. The drum used in the test contains a shelf which lifts and drops the aggregate sample and steel balls during a specified number of revolutions. The tumbling action, combined with impact causes the more brittle particles to shatter. The aggregate sample is removed, screened, and measured losses based on the percentage of material passing the specified screen sizes are determined.

### MAGNESIUM SULFATE SOUNDNESS LOSS:

The magnesium sulfate soundness loss test is determined by the ASTM C 88 test method. The sulfate soundness loss test is intended to provide an estimate of the resistance of aggregate to the weathering action that occurs in concrete. The measured and screened aggregate sample is placed in a magnesium sulfate salt solution for a specified period of time. The sample is removed, dried and re-immersed for five cycles. The sample is then removed, screened and weighed to determine the percentage loss of material by screen size. A final weighted average loss for each size fraction is calculated.

*Figure 2. Summary of important quality tests performed on aggregate materials from (Marek, 1991).*

### REFERENCES CITED


MINERALOGY, GEOTHERMOMETRY, AND GEOBAROMETRY OF THE ROWLAND SPRING FORMATION, ALLATOONA COMPLEX, GEORGIA

RANDY L. KATH
Department of Geology
University of West Georgia
Carrollton, Georgia 30118-3100

THOMAS J. CRAWFORD
Department of Geology
University of West Georgia
Carrollton, Georgia 30118-3100

ABSTRACT

In the Allatoona Dam 7.5 minute quadrangle of northern Georgia, the Rowland Spring Formation occurs as xenoliths and roof pendants in the Corbin Metagranite. The Rowland Spring Formation was metamorphosed to pyroxene-granulite facies before being intruded by the Corbin Metagranite, which is younger than the pyroxene-granulite facies metamorphism.

The Rowland Spring Formation is divided into two geochemically and petrologically distinct groups of lithologies: a fine-grained, Mg-rich group of lithologies; and a coarser-grained, Fe-rich group of lithologies. In general, the Rowland Spring Formation is characterized by four equilibrium mineral assemblages:

1) quartz + biotite + garnet + plagioclase + K-feldspar
2) quartz + garnet + plagioclase + K-feldspar
3) quartz + orthopyroxene + garnet + plagioclase + K-feldspar
4) quartz + orthopyroxene + clinopyroxene + garnet + plagioclase

The orthopyroxene is typically coarse-grained and subhedral; however, orthopyroxene exsolution lamellae occur in a clinopyroxene host. Locally, orthopyroxene is altered to chlorite, indicating retrograde metamorphic overprinting of a primary pyroxene. The coexistence of augite and orthopyroxene is indicative of inverted pigeonite, suggesting temperatures in excess of 825 °C. Garnet-biotite core temperatures, and garnet-clinopyroxene geothermometry indicate temperatures between 680 °C and 925 °C, respectively. Garnet-orthopyroxene geobarometry indicates pressures around 12 kbars to 13 kbars. Based on these temperatures and pressures, we estimate that the Rowland Spring Formation was metamorphosed at a depth of 30 km to 35 km under fairly dry conditions. These data provide some of the first quantitative evidence for granulite facies metamorphism in the southernmost Blue Ridge Physiographic Province in Georgia.

INTRODUCTION

The Cartersville District has been the subject of intense study since the first published papers by Hayes in the late 1800’s. Because of its structural complexity and economic potential, many works have been published on this area since Hayes’ first work in 1891 (Hayes and Eckel, 1902; Watson, 1902; Crickmay, 1936; Kesler, 1950; Croft, 1963; Morgan, 1966; Odom and others, 1973; Martin, 1974; Cressler and Crawford, 1976 in Cressler and others, 1979; Costello, 1978; McConnell and Costello, 1980; Gargi, 1985; Higgins and others, 1996; Kath and
KATH AND CRAWFORD

others, 1996; Heatherington and others, 1996). Many of these papers focused on the origin and tectonostratigraphic position of the Corbin Metagranite (referred to as the Corbin) and its significance to the Grenville Orogeny (Higgins and others, 1989; Heatherington and others, 1996). Based on age dates by Odom and others, (1973) and Heatherington and others, (1996), the Corbin is thought to have intruded between 1.1 Ga and 1.5 Ga. These ages represent the oldest known age-dated lithologies in Georgia. Although most of the research in this area has concentrated on the Corbin, few papers have concentrated on the metamorphic history of the Corbin or associated lithologies with the exception of Martin (1974) and Kath and others, (1996). Martin’s work focused on the petrogenesis and metamorphism of the Corbin Metagranite, but did not address lithologies associated with the Corbin Metagranite.

Detailed geologic mapping of the Allatoona Dam Quadrangle by Cressler and Crawford (1976), Crawford and others, (2009), Crawford and others, (2010), and Kath and others (2015, Plate 1, this volume) have shown the presence of large masses of distinctly different lithologies in the Corbin Metagranite (Figure 1 and Plate 1). These lithologies occur as roof pendants and xenoliths within the main porphyritic granite phase of the Corbin. Based on their relationship with the Corbin Metagranite, these other lithologic units must represent older lithologies which were intruded by the granite, indicating that the Corbin is not the oldest rock in Georgia.

There is a distinct lack of quantitative data pertaining to the metamorphic history of the Cartersville District. Because of fresh exposures of roof pendant and xenolith lithologies in Vulcan’s Bartow County Quarry (DeCinque and Nelson, 1996 and this volume), the authors have been able to determine preliminary mineral assemblages, geothermometry, and geobarometry from lithologies intruded by the Corbin. These older rocks have been named the Rowland Spring Formation by Crawford and others, (1999).

GENERAL DESCRIPTION

Outcrops of the Rowland Spring Formation occur throughout the Allatoona Dam, Georgia 7.5-minute quadrangle (Figure 1 and Plate 1). Fresh samples were taken from xenoliths and/or roof pendants in the interior of the outcrop of the Corbin and near the margin, along the entrance road to Vulcan’s Bartow County quarry. All samples were thin-sectioned and described petrographically. From both field and petrographic descriptions, the Rowland Spring Formation was classified based on textural and mineralogic parameters.

The Rowland Spring Formation contains at least two distinct lithologies: one is a coarse-grained phase and the other is a fine-grained phase. The coarse-grained phase is the dominant and is characterized as a light-colored, relatively coarse-grained, porphyroblastic-garnet, biotite, quartz, feldspar, granofelsic-gneiss with minor crystalline graphite and sulfides. The fine-grained phase occurs within this coarse-grained phase and is characterized as a thin, dark-colored, weakly-zoned, orthopyroxene, ± clinopyroxene, quartz, ± garnet, feldspar, granofels. The dark-colored granofels phases are lenticular and generally oriented with their long axes parallel to the plane of foliation and compositional layering. Mineral assemblages observed in the Rowland Spring Formation are listed on Table 1.

ANALYTICAL PROCEDURES

Mineral Identification and Analysis

Samples were examined in thin section and mineral identification was aided by use of a JEOL 8900 energy dispersive system on the electron microprobe. Selected samples were analyzed with a four-spectrometer JEOL 8900 fully automated microprobe using natural and synthetic silicate and oxide standards. Standard
Figure 1. Geologic and sample location map of the Allatoona Dam, Georgia, 7.5 minute quadrangle. Original geologic mapping by Cressler and Crawford (1976) in Cressler and others, (1979, Plate 4); expanded and modified by Crawford and others, 2009; Crawford and others, 2010; and Plate 1 (this volume). See Plate 1 for a detailed lithologic and structural legend.
operating conditions for the silicates are 15 keV accelerating potential and 1 µm beam spot. Data were reduced on-line using a digital computer. Precision was monitored by analyzing standards and is ±2 % for the major elements and ±5% for the minor elements. Unknown minerals were analyzed to determine intra- and inter-granular homogeneities. Several areas within each thin section were analyzed to determine compositional ranges. Coexisting phases used in equilibrium calculations were analyzed in 2.5 mm domains within each sample. For mineral rim analyses, the electron beam was focused approximately 5 µm to 10 µm from the rim to avoid any edge effects.

Each sample was trimmed to the dimensions of a thin section (22mm x 46mm) and then sawed to yield a thin section chip and a sample which was powdered for wet chemistry and x-ray diffraction. This procedure was followed to ensure that the thin section and powder correspond as much as possible to the same part of the rock.

**MINERAL CHEMISTRY**

**Biotite**

Biotite is a nearly ubiquitous phase in both the coarse- and fine-grained Rowland Spring Formation. The major compositional variation within biotite is in the Fe/Mg and Ti content (Table 2). In the coarse-grained Rowland Spring Formation, Fe ranges from 0.554 to 1.044 cations per 11 oxygens. Ti concentrations in biotite range from 0.110 to 0.395 cations per 11 oxygens. The variation in Ti concentration appears to be a function of the saturating effects of ilmenite; when biotite directly coexists with ilmenite, the biotite contains less Ti.

**TABLE I**

<table>
<thead>
<tr>
<th>Sample #</th>
<th>Fm</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Bio</th>
<th>Musc</th>
<th>Gar</th>
<th>Ca-Pyx</th>
<th>Fe-Pyx</th>
<th>K-sp</th>
<th>Plg</th>
<th>Qtz</th>
<th>Ilm</th>
<th>Mag</th>
<th>Ap</th>
<th>Zr</th>
<th>Sph</th>
<th>Mon</th>
</tr>
</thead>
<tbody>
<tr>
<td>95-AC-01</td>
<td>RS-C</td>
<td>34°07'12&quot;</td>
<td>84°43'44&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-02</td>
<td>RS-C</td>
<td>34°07'31&quot;</td>
<td>84°43'07&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-03</td>
<td>RS-C</td>
<td>34°06'54&quot;</td>
<td>84°41'23&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-04</td>
<td>RS-C</td>
<td>34°12'59&quot;</td>
<td>84°43'48&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-05</td>
<td>RS-C</td>
<td>34°12'59&quot;</td>
<td>84°43'48&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-06</td>
<td>RS-C</td>
<td>34°12'59&quot;</td>
<td>84°43'48&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-07</td>
<td>RS-C</td>
<td>34°12'59&quot;</td>
<td>84°43'48&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-08</td>
<td>RS-C</td>
<td>34°12'59&quot;</td>
<td>84°43'48&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-09</td>
<td>RS-C</td>
<td>34°12'59&quot;</td>
<td>84°43'48&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-10</td>
<td>RS-C</td>
<td>34°12'59&quot;</td>
<td>84°43'48&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-11</td>
<td>RS-C</td>
<td>34°12'59&quot;</td>
<td>84°43'48&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-12</td>
<td>RS-C</td>
<td>34°12'59&quot;</td>
<td>84°43'48&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-13</td>
<td>RS-C</td>
<td>34°12'59&quot;</td>
<td>84°43'48&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>95-AD-14</td>
<td>RS-C</td>
<td>34°12'59&quot;</td>
<td>84°43'48&quot;</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td>✔</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Notes: AD = Allatoona Dam Quadrangle  
AC = Acworth Quadrangle  
RS = Rowland Spring Formation  
-C = Coarse-grained portion of the Rowland Spring Formation  
-F = Fine-grained portion of the Rowland Spring Formation  
Bio = biotite, Musc = muscovite, Gar = garnet, Ca-Pyx = Ca-rich clinopyroxene, Fe-Pyx = Fe-rich orthopyroxene, K-sp = K-feldspar, Plg = plagioclase, Qtz = quartz, Ilm = ilmenite, Mag = magnetite, Ap = apatite, Zr = zircon, Sph = sphene, Mon = monozite  
¹ Electron Microprobe Data
MINERALOGY, GEOTHERMOMETRY, AND GEOBAROMETRY OF THE ROWLAND SPRING FORMATION

Garnet

Garnets are dominantly almandine-rich, but display variations in Mn, Ca, and Mg. The compositions of garnets in the Rowland Spring Formation (Table 3) reflect the high total iron or magnesium content of the rock. All analyzed garnets are generally stoichiometric \((\text{Fe,Mn,Mg,Ca})_3\text{Al}_2\text{Si}_3\text{O}_{12}\). \(\text{Cr}_2\text{O}_3\) is present in most of the garnets although it is generally a minor component.

Compositional zoning of garnet in the Rowland Spring Formation is virtually nonexistent except for minor variations along the extreme edge (rim) of the garnet. Garnet zoning profiles from the Rowland Spring Formation, plotted on Figure 4, illustrate that there is virtually no compositional zoning in the spessartine or grossular components. The lack of a zonation pattern is consistent for garnets within a particular mineral assemblage.

Pyroxene

Pyroxenes are dominantly Fe- and Mg-rich, but display significant variation in Ca and Mn. All analyzed pyroxenes are generally stoichiometric \((\text{Fe,Mn,Mg,Ca})_2\text{Si}_2\text{O}_6\). \(\text{Cr}_2\text{O}_3\) is present in most of the pyroxenes although it is generally a minor component.
Two pyroxene compositions are common in the Rowland Spring Formation: a Ca-rich clinopyroxene and a Fe-rich orthopyroxene. The compositions of the pyroxenes are shown on a pyroxene quadrilateral, presented as Figure 5. Based on Figure 5, the Ca-rich clinopyroxene plots nearly equidistant between diopside and hedenbergite. For the purpose of this paper, the Ca-rich clinopyroxene is classified as a ferroaugite. The Fe-rich orthopyroxene is slightly skewed toward the ferrosillite end member of the pyroxene quadrilateral and is therefore classified as a hypersthene.

The two pyroxenes in the Rowland Spring Formation occur as individual ferroaugite and hypersthene grains as well as exsolution lamellae. The pyroxene grains which show exsolution lamellae are characterized by a ferroaugite host with hypersthene lamellae. A computer-enhanced, backscatter photomicrograph from sample 95AD11 is shown on Figure 6. The computer enhancement was performed to show the fine hypersthene lamellae in the ferroaugite host.

MINERAL ASSEMBLAGES

Phases of the Rowland Spring Formation can be represented in the system: SiO$_2$-Al$_2$O$_3$-Cr$_2$O$_3$-TiO$_2$-FeO-MnO-MgO-K$_2$O-H$_2$O-CO$_2$. Several assumptions must be made in order to reduce the number of components for graphical representation. For example, TiO$_2$ is assumed to be present only in biotite and ilmenite which are ubiquitous phases to both high- and low-variance assemblages. Quartz is present in all assemblages; therefore, SiO$_2$ is considered to be in excess. Biotite and K-feldspar are the only K-bearing phases and are nearly ubiquitous to both high- and low-variance assemblages; therefore, K$_2$O must be considered as a component. Cr$_2$O$_3$ occurs only in trace quantities in biotite, garnet, and pyroxene; and Mn is a minor component in only garnet and pyroxene; therefore, these components are not
### TABLE 2
Representative Biotite Analyses from the Rowland Spring Formation

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>95AD02</th>
<th>95AD14</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>39.18</td>
<td>39.70</td>
</tr>
<tr>
<td>TiO₂</td>
<td>2.87</td>
<td>4.12</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>0.06</td>
<td>0.03</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>14.63</td>
<td>13.80</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.59</td>
<td>0.57</td>
</tr>
<tr>
<td>MnO</td>
<td>0.59</td>
<td>0.57</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>15.93</td>
<td>16.33</td>
</tr>
<tr>
<td>CaO</td>
<td>3.59</td>
<td>3.79</td>
</tr>
<tr>
<td>MgO</td>
<td>14.52</td>
<td>12.72</td>
</tr>
<tr>
<td>K₂O</td>
<td>7.37</td>
<td>4.99</td>
</tr>
<tr>
<td>Total</td>
<td>97.68</td>
<td>96.20</td>
</tr>
</tbody>
</table>

### TABLE 3
Representative Garnet Analyses from the Rowland Spring Formation

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>95AD02</th>
<th>95AD14</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>38.96</td>
<td>39.11</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.03</td>
<td>0.05</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>22.81</td>
<td>24.21</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>5.99</td>
<td>5.57</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.02</td>
<td>0.01</td>
</tr>
<tr>
<td>MgO</td>
<td>6.40</td>
<td>6.20</td>
</tr>
<tr>
<td>CaO</td>
<td>1.61</td>
<td>1.50</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td>Total</td>
<td>97.61</td>
<td>97.83</td>
</tr>
</tbody>
</table>

Note: All Fe calculated as FeO

---

**MINERALOGY, GEOTHERMOMETRY, AND GEOBAROMETRY OF THE ROWLAND SPRING FORMATION**

Cations based on 11 oxygens:

- **Si**: 2.868, 2.845, 2.821, 2.897, 2.837, 2.822, 2.788, 2.799, 2.799, 2.815, 2.936, 2.846
- **Al**°: 1.141, 1.155, 1.179, 1.103, 1.163, 1.178, 1.212, 1.201, 1.201, 1.185, 1.064, 1.154
- **Mg**: 0.159, 0.237, 0.164, 0.168, 0.222, 0.110, 0.287, 0.255, 0.282, 0.226, 0.177, 0.215
- **Cr**: 0.004, 0.002, 0.002, 0.005, 0.005, 0.002, 0.004, 0.006, 0.008, 0.004, 0.023, 0.005
- **Ti**: 1.270, 1.244, 1.241, 1.261, 1.205, 1.218, 1.278, 1.267, 1.270, 1.295, 1.497, 1.323
- **Fe**: 0.156, 0.089, 0.062, 0.159, 0.042, 0.040, 0.066, 0.085, 0.069, 0.110, 0.433, 0.168
- **Fe³⁺**: 0.821, 1.044, 0.932, 0.747, 0.974, 0.858, 0.735, 0.761, 0.762, 0.777, 0.759, 0.606
- **Mg**: 0.593, 0.450, 0.619, 0.570, 0.570, 0.495, 0.526, 0.512, 0.506, 0.528, 0.485, 0.364
- **Ca**: 0.000, 0.000, 0.000, 0.000, 0.000, 0.000, 0.000, 0.000, 0.000, 0.000, 0.000, 0.000
- **K**: 0.914, 0.025, 0.931, 0.893, 0.902, 0.884, 0.898, 0.892, 0.848, 0.840, 0.673, 0.864
- **Fe³⁺**: 0.926, 0.944, 0.943, 0.902, 0.912, 0.906, 0.915, 0.915, 0.862, 0.867, 0.741, 0.890
- **XFe**: 0.387, 0.418, 0.436, 0.386, 0.329, 0.262, 0.298, 0.284, 0.306, 0.306, 0.308, 0.315
- **XTi**: 0.056, 0.084, 0.056, 0.059, 0.077, 0.036, 0.102, 0.089, 0.099, 0.079, 0.065, 0.076

Note: All Fe calculated as FeO

XFe = n \* Fe(nFe + nMg + nMn + nTi)

---

**TABLE 2**

**TABLE 3**

**Note**: All Fe calculated as FeO

XFe = n \* Fe(nFe + nMg + nMn + nCa)
considered to be significant. Both H2O and CO2 are considered boundary value components (Zen, 1967). The resulting simplified subsystem is Al2O3-FeO-MgO-CaO-K2O (for original assumptions see Thompson, 1957).

The resulting subsystem would require three or four dimensions to be properly represented. Because biotite and K-feldspar are the only K-bearing phases, the components can be further reduced to two subsystems, Al2O3-CaO-FeO-K2O (ACFK) and Al2O3-FeO-MgO-K2O (AFMK).

Because K-feldspar is present in all assemblages, both the ACFK and AFMK subsystems were projected from measured K-feldspar compositions to create AFM and ACF diagrams. Figure 7 represents the metamorphic facies types for the Rowland Spring Formation constructed from the known mineral assemblages and mineral chemistries (Tables 1 through 3). For any given assemblage, the tie lines on the AFM and ACF diagrams do not cross. Any inconsistencies are attributed to changes in temperature, minor changes in µH2O, or stabilizing effects of additional components. Mineral assemblages which contain pyroxene exsolution textures show the most inconsistencies.

**Garnet-Biotite Assemblage**

The assemblages quartz + biotite + garnet + plagioclase + K-feldspar, and quartz + garnet + plagioclase + K-feldspar are observed in the coarse-grained, garnet-rich portions of the Rowland Spring Formation. The AFM

---

**Table 4**

Representative Pyroxene Analyses from the Rowland Spring Formation

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>9SAD11</th>
<th>SiO2</th>
<th>TiO2</th>
<th>Cr2O3</th>
<th>Al2O3</th>
<th>FeO</th>
<th>MnO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na2O</th>
<th>K2O</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>9SAD11</td>
<td></td>
<td>50.98</td>
<td>0.12</td>
<td>0.11</td>
<td>1.94</td>
<td>29.05</td>
<td>1.79</td>
<td>15.93</td>
<td>0.38</td>
<td>0.03</td>
<td>0.05</td>
<td>100.36</td>
</tr>
<tr>
<td>9SAD11</td>
<td></td>
<td>51.06</td>
<td>0.10</td>
<td>0.11</td>
<td>1.79</td>
<td>29.07</td>
<td>1.78</td>
<td>16.14</td>
<td>0.35</td>
<td>0.01</td>
<td>0.02</td>
<td>100.40</td>
</tr>
<tr>
<td>9SAD11</td>
<td></td>
<td>51.22</td>
<td>0.13</td>
<td>0.05</td>
<td>0.86</td>
<td>29.09</td>
<td>2.27</td>
<td>15.62</td>
<td>0.79</td>
<td>0.01</td>
<td>0.02</td>
<td>100.05</td>
</tr>
<tr>
<td>9SAD11</td>
<td></td>
<td>51.40</td>
<td>0.14</td>
<td>0.08</td>
<td>0.99</td>
<td>29.21</td>
<td>2.30</td>
<td>15.75</td>
<td>0.74</td>
<td>0.01</td>
<td>0.02</td>
<td>99.97</td>
</tr>
<tr>
<td>9SAD11</td>
<td></td>
<td>51.18</td>
<td>0.09</td>
<td>0.00</td>
<td>0.65</td>
<td>29.35</td>
<td>1.64</td>
<td>15.67</td>
<td>0.84</td>
<td>0.10</td>
<td>0.02</td>
<td>100.54</td>
</tr>
<tr>
<td>9SAD11</td>
<td></td>
<td>51.60</td>
<td>0.14</td>
<td>0.00</td>
<td>0.77</td>
<td>28.77</td>
<td>2.41</td>
<td>16.03</td>
<td>0.78</td>
<td>0.01</td>
<td>0.01</td>
<td>100.54</td>
</tr>
<tr>
<td>9SAD11</td>
<td></td>
<td>51.36</td>
<td>0.10</td>
<td>0.00</td>
<td>0.59</td>
<td>28.51</td>
<td>2.45</td>
<td>15.77</td>
<td>0.79</td>
<td>0.01</td>
<td>0.01</td>
<td>100.33</td>
</tr>
<tr>
<td>9SAD11</td>
<td></td>
<td>51.36</td>
<td>0.11</td>
<td>0.00</td>
<td>0.58</td>
<td>29.35</td>
<td>2.38</td>
<td>15.50</td>
<td>0.96</td>
<td>0.01</td>
<td>0.01</td>
<td>100.69</td>
</tr>
<tr>
<td>9SAD11</td>
<td></td>
<td>51.45</td>
<td>0.09</td>
<td>0.04</td>
<td>0.58</td>
<td>11.98</td>
<td>1.07</td>
<td>15.84</td>
<td>0.93</td>
<td>0.01</td>
<td>0.03</td>
<td>100.97</td>
</tr>
<tr>
<td>9SAD11</td>
<td></td>
<td>51.22</td>
<td>0.21</td>
<td>0.00</td>
<td>1.20</td>
<td>11.96</td>
<td>1.13</td>
<td>15.46</td>
<td>21.68</td>
<td>0.13</td>
<td>0.03</td>
<td>99.69</td>
</tr>
<tr>
<td>9SAD11</td>
<td></td>
<td>51.86</td>
<td>0.25</td>
<td>0.01</td>
<td>1.33</td>
<td>11.96</td>
<td>1.13</td>
<td>15.36</td>
<td>21.62</td>
<td>0.15</td>
<td>0.03</td>
<td>98.81</td>
</tr>
</tbody>
</table>

Note: All Fe calculated as FeO

Xi = n i/(nFe + nMg + nMn + nCa)
Figure 7. Al2O3-CaO-FeO-MgO Mineral Assemblages from the Rowland Spring Formation.
assemblage or facies is shown on Figure 7A. For any given assemblage, the tie lines do not cross. Additionally, the assemblage shown on Figure 7A represent some of the most Fe-rich garnet compositions observed in the Rowland Spring Formation. This assemblage was observed in samples taken from Vulcan’s Bartow County quarry road (95AD14) and from a small xenolith (95AD02) in the central part of the quadrangle (see Figure 1).

Garnet-Biotite-Pyroxene Assemblage
The assemblages quartz + orthopyroxene + garnet + plagioclase + K-feldspar and quartz + orthopyroxene + clinopyroxene + garnet + plagioclase are restricted to the finer-grained portions of the Rowland Spring Formation. The ACF assemblage or facies is shown on Figure 7B and 7C. For any given assemblage, the tie lines do not cross. Additionally, the assemblage shown on Figure 7B and 7C represent the most Mg-rich garnet compositions observed in the Rowland Spring Formation. This assemblage was observed in samples taken from Vulcan’s Bartow County quarry road (95AD11).

INTENSIVE PARAMETERS
A fundamental goal in metamorphic petrology is to quantify the physical and chemical conditions in the crust during metamorphism. With the introduction of thermodynamics to petrology, quantification of the intensive parameters during metamorphism can be achieved. During prograde and retrograde metamorphism it is useful to estimate the T and P at which the mineral assemblages equilibrated. Considering the mineral assemblages present in the Rowland Spring Formation, the two most effective geothermometers are garnet-biotite and garnet-pyroxene and the most effective geobarometer is garnet-pyroxene-plagioclase. These are discussed in the following sections.

Temperature of Metamorphism
Garnet-Biotite Geothermometry-
Iron and Mg partitioning between coexisting garnet and biotite is temperature dependent and therefore can be used as a geothermometer (Ferry and Spear, 1978; Thompson, 1976; Goldman and Albee, 1977; Ghent and Stout, 1981; Hodges and Spear, 1982; Ganguly and Saxena, 1984). This geothermometer is based on the following exchange reaction:

\[
\text{KFe}_3\text{Al}(\text{Si}_3\text{Al})\text{O}_{10}(\text{OH})_2 + \text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12} = \text{KFe}_3\text{Al}(\text{Si}_3\text{Al})\text{O}_{10}(\text{OH})_2 + \text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12} \quad (A)
\]

The exchange operator is FeMg\textsuperscript{1}. Because the exchange reaction is independent of fluid composition, and the distribution coefficient

\[
K_d = [\text{XMg}/\text{XFe}]_{\text{Ga}}/[\text{XMg}/\text{XFe}]_{\text{Bio}}
\]  

is only weakly dependent on pressure (Ferry and Spear, 1978), the partitioning is dependent only on temperature.

The garnet-biotite exchange equilibria have been experimentally calibrated by Ferry and Spear (1978) and Perchuk and Laurent'eva (1983). The calibration of Ferry and Spear (1978) has been modified for the non-ideal mixing behavior of Ca and Mn by Ganguly (1979), Hodges and Spear (1982), Ganguly and Saxena (1984), and Hoinkes (1986); and adjustments for Ti and ViAl in biotite were discussed by Indares and Martignole (1985). The equilibrium temperatures recorded by the coexisting minerals were calculated based on the following relationships for \( K_d \) and T:

\[
T(K) = \frac{(4151 - 0.8W_{\text{FeMg}} + 18.8P_{\text{kbars}})}{(1.987 + \ln K_d + 1.554)} + \frac{[W_{\text{FeMg}}(X_{\text{Fe}} - X_{\text{Mg}})]^{\text{B}} + \Delta W_{\text{gr}} * X_{\text{gr}} + \Delta W_{\text{sp}} * X_{\text{sp}}}{(1.987 + \ln K_d + 1.554)}
\]  

based on Ganguly and Saxena (1984) where:

\[
W_{\text{FeMg}} = [W_{\text{FeMg}}(\text{Mg}/(\text{Mg+Fe})) + W_{\text{Mg}} * F_{\text{Fe}}/(\text{Fe} + \text{Mg})]^{\text{B}}
\]
\[ W_{Mg-Fe} = 2500 \pm 500 \text{ cal mol}^{-1} \]
\[ W_{Fe-Mg} = 200 \pm 500 \text{ cal mol}^{-1} \]
\[ \Delta W^{HF} = \Delta W^{pp} = 3000 \text{ cal mol}^{-1} \]

\[
T(K) = \frac{(-2109 + 0.155P(\text{bars}))}{(-\ln K_d - 0.782)}
\]

(3)

based on polybaric equation 7 presented by Ferry and Spear (1978); and

\[
T(K) = \frac{(12454 - 0.057P(\text{bars}) + 3\alpha_{\text{Ca}a}^{\text{Ga}} + 3\alpha_{\text{Fe}a}^{\text{Ga}} - \Delta W^{HF}_{\text{Ga}} - \Delta W^{HF}_{\text{Mg}})}{(4.662 - 5.966\ln K_d)}
\]

(4)

based on Indares and Martignole (1985).

**Garnet-Clinopyroxene Geothermometry-**

Iron and Mg partitioning between coexisting garnet and clinopyroxene has also been recognized to be temperature dependent (Saxena, 1968; Banno, 1970; Myesn and Heier, 1972; Ganguly, 1979). This geothermometer is based on the following exchange reaction:

\[
\begin{align*}
\text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12} + 3\text{CaFeSi}_2\text{O}_6 & = \\
(\text{garnet}) & (\text{clinopyroxene}) \\
\text{Fe}_3\text{Al}_2\text{Si}_3\text{O}_{12} + 3\text{CaMgSi}_2\text{O}_6 & = \\
(\text{garnet}) & (\text{clinopyroxene})
\end{align*}
\]

(B)

The exchange operator is \( \text{FeMg}^{-1} \). Because the reaction exchange is independent of fluid composition, and the distribution coefficient

\[ K_d = \frac{[X_{Fe}/XMg]^G}{[X_{Fe}/XMg]^C} \]

is only weakly dependent on pressure (Ganguly, 1979), the partitioning is dependent only on temperature.

The garnet-clinopyroxene exchange equilibria have been calibrated by Ganguly (1979). The calibration of Ganguly (1979) has been modified for the non-ideal mixing behavior of Ca and Mg between the M1- and M2-sites by Newton (1983). The equilibrium temperatures recorded by the coexisting minerals were calculated based on the following relationships for \( K_d \) and \( T \) (equation 10b of Ganguly (1979):

\[
T(K) = \frac{(4801 + 11.07P)}{(\ln(K_d) + 2.930 - \ln(\psi))}
\]

(6)

Because of the lack of adequate mixing data on the diopside and hedenbergite components, Ganguly (1979) restricted his equations to Na-poor compositions so that \( \ln(\psi) \) could be neglected.

**Pressure of Metamorphism**

**Garnet-Orthopyroxene-Plagioclase Geobarometry**

Several mineralogic geobarometers for metamorphic rocks have been proposed which make use of the common coexistence of garnet with aluminous minerals such as muscovite and plagioclase (Gent and Stout, 1981; Newton and Perkins, 1982). In granulite facies metamorphic rocks, muscovite is not a stable phase of an equilibrium mineral assemblage; therefore, the geobarometric equations must exclude muscovite. Newton and Perkins (1982) have introduced a mineralogic geobarometer which is applicable to mineralogic assemblages common in granulite-grade quartzofeldspathic and basic lithologies. The geobarometer is based on the following exchange reaction:

\[
\begin{align*}
\text{CaAl}_2\text{Si}_2\text{O}_8 + \text{Mg}_2\text{Si}_2\text{O}_6 & = \\
(\text{plagioclase}) & (\text{opx}) \\
1/3\text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12} + 2/3\text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12} + \text{SiO}_2 & = \\
(\text{garnet}) & (\text{quartz})
\end{align*}
\]

(C)

\[ P_{\text{opx}}(\text{bars}) = 3944 + 13.070T(K) + 3.5038\ln(K_d) \]

(7)

where

\[ K_d = \frac{(a_{\text{Ca}}a_{\text{Mg}}^{\text{Ga}})^{\text{Ga}}}{(a_{\text{Ca}}a_{\text{Mg}}^{\text{Op})}^{\text{Op}}} \]

(8)

Based on calorimetric uncertainties in \( \Delta G_A^{\circ} \), the pressure uncertainty is ±1500 bars (Newton and Perkins, 1982). Additionally, Newton and Perkins present a geobarometric expression for clinopyroxene-bearing assemblages; however, they note that pressures calculated using clinopyroxene are several
### TABLE 5

**Garnet-Biotite and Biotite-Clinopyroxene Geothermometry**

from the Rowland Spring Formation, Allatoona Complex, Georgia

<table>
<thead>
<tr>
<th>Sample</th>
<th>Pair</th>
<th>Garnet</th>
<th>Biotite</th>
<th>Clinopyroxene</th>
<th>Location</th>
<th>Kd</th>
<th>Temperature (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>95AD11</td>
<td>02-19</td>
<td>0.571</td>
<td>0.067</td>
<td>0.046 0.316</td>
<td>na</td>
<td>3.079</td>
<td>946</td>
</tr>
<tr>
<td>95AD11</td>
<td>13-19</td>
<td>0.596</td>
<td>0.071</td>
<td>0.060 0.272</td>
<td>na</td>
<td>3.723</td>
<td>885</td>
</tr>
<tr>
<td>95AD11</td>
<td>22-19</td>
<td>0.582</td>
<td>0.068</td>
<td>0.055 0.296</td>
<td>na</td>
<td>3.359</td>
<td>921</td>
</tr>
<tr>
<td>95AD11</td>
<td>02-19</td>
<td>0.571</td>
<td>0.067</td>
<td>0.046 0.316</td>
<td>na</td>
<td>3.060</td>
<td>948</td>
</tr>
<tr>
<td>95AD11</td>
<td>13-19</td>
<td>0.596</td>
<td>0.071</td>
<td>0.060 0.272</td>
<td>na</td>
<td>3.700</td>
<td>883</td>
</tr>
<tr>
<td>95AD11</td>
<td>22-19</td>
<td>0.582</td>
<td>0.068</td>
<td>0.055 0.296</td>
<td>na</td>
<td>3.274</td>
<td>928</td>
</tr>
<tr>
<td>95AD11</td>
<td>02-22</td>
<td>0.571</td>
<td>0.067</td>
<td>0.046 0.316</td>
<td>na</td>
<td>3.001</td>
<td>954</td>
</tr>
<tr>
<td>95AD11</td>
<td>13-22</td>
<td>0.596</td>
<td>0.071</td>
<td>0.060 0.272</td>
<td>na</td>
<td>3.629</td>
<td>899</td>
</tr>
<tr>
<td>95AD11</td>
<td>22-22</td>
<td>0.582</td>
<td>0.068</td>
<td>0.055 0.296</td>
<td>na</td>
<td>3.274</td>
<td>928</td>
</tr>
</tbody>
</table>

**Geometric Mean** 95AD11: 895

**Standard Deviation:** 23


---

### TABLE 6

**Garnet-Plagioclase-Orthopyroxene Geobarometry**

from the Rowland Spring Formation, Allatoona Complex, Georgia

<table>
<thead>
<tr>
<th>Sample</th>
<th>Pair</th>
<th>Garnet</th>
<th>Orthopyroxene</th>
<th>Plagioclase</th>
<th>Kd</th>
<th>Pressure (bars)</th>
</tr>
</thead>
<tbody>
<tr>
<td>95AD11</td>
<td>02-18</td>
<td>0.571</td>
<td>0.067 0.467</td>
<td>0.406 0.467</td>
<td>0.208</td>
<td>0.459 0.300 0.199</td>
</tr>
<tr>
<td>95AD11</td>
<td>13-19</td>
<td>0.596</td>
<td>0.071 0.467</td>
<td>0.406 0.467</td>
<td>0.208</td>
<td>0.459 0.300 0.199</td>
</tr>
<tr>
<td>95AD11</td>
<td>22-19</td>
<td>0.582</td>
<td>0.068 0.467</td>
<td>0.406 0.467</td>
<td>0.208</td>
<td>0.459 0.300 0.199</td>
</tr>
<tr>
<td>95AD11</td>
<td>02-22</td>
<td>0.571</td>
<td>0.067 0.467</td>
<td>0.406 0.467</td>
<td>0.208</td>
<td>0.459 0.300 0.199</td>
</tr>
<tr>
<td>95AD11</td>
<td>13-22</td>
<td>0.596</td>
<td>0.071 0.467</td>
<td>0.406 0.467</td>
<td>0.208</td>
<td>0.459 0.300 0.199</td>
</tr>
<tr>
<td>95AD11</td>
<td>22-22</td>
<td>0.582</td>
<td>0.068 0.467</td>
<td>0.406 0.467</td>
<td>0.208</td>
<td>0.459 0.300 0.199</td>
</tr>
</tbody>
</table>

**Geometric Mean (kbar):** 12.92

**Standard Deviation (kbar):** 0.59

kilobars lower than the orthopyroxene-bearing assemblage.

**DISCUSSION**

Temperatures were calculated using garnet-biotite geothermometry for samples 95AD02 and 95AD14 (Table 5). Using temperatures calculated for both core- and rim-chemistry in the garnets, temperatures ranged between 558 °C and 760 °C using the non-ideal mixing model of Ganguly and Saxena (1984). Temperatures calculated using Ganguly and Saxena are generally 40 °C to 90 °C higher than those calculated using the ideal mixing model of Ferry and Spear (1978). The differences in the calculated temperatures are due to the Mn and Ca components of the garnets. Because the garnets in the Rowland Spring Formation are relatively low in Mn (XMn < 0.050) and moderately low in Ca (XCa < 0.145) both methods of calculation should give reasonably similar temperatures. The differences in calculated temperatures suggest that the activity coefficients and activity-composition relations in the garnet quaternary system are still poorly understood (Koziol and Newton, 1989; Thoenen, 1986). Because of the grossular component in analyzed garnets, we prefer to use the non-ideal mixing model of Ganguly and Saxena (1984).

Calculated temperatures for garnet-biotite pairs using garnet rim- and core-data are shown on Table 5. In general, temperatures calculated using garnet core-data are up to 128 °C hotter than those calculated for rim-data, Table 5. Garnet core temperatures from sample 95AD14 and 95AD02 are estimated to be 748 °C and 688 °C, respectively. In comparison, garnet rim temperatures from the same two samples are estimated to be 644 °C and 560 °C, respectively. Although there is very little compositional zoning in the garnets (Figure 4) the zoning that is present is seen near the extreme rim portion of the garnet. In general, the rim composition is higher in XFe and lower in XMg, suggesting a possible re-equilibration during superposed metamorphic events or during the waning stages of the prograde metamorphic event.

Garnet-clinopyroxene temperatures calculated for sample 95AD11 using the calibration of Ganguly (1979) are presented on Table 5. Temperatures ranged between 892 °C and 954 °C, with an average of 923 °C and a standard deviation of 23 °C. The range in garnet-clinopyroxene temperatures from this sample is relatively narrow indicating that both garnet and clinopyroxene were in equilibrium and that the garnets were virtually unzoned.

The lower temperatures obtained from the garnet-biotite geothermometry suggest that: 1) garnet-biotite pairs record different peak metamorphic conditions due to lower equilibration temperature; 2) the garnet-biotite pairs re-equilibrated during a second phase of metamorphism which was lower than the peak metamorphic conditions recorded by pyroxene equilibria; or 3) the large bulk compositional differences between the fine-grained and coarse-grained Rowland Spring Formation preclude exact comparison of calculated temperatures, because the mixing parameters for Ca and Mn in garnet and/or pyroxene are poorly known.

During cooling, natural pyroxenes undergo two changes: 1) intracrystalline exsolution; or 2) intercrystalline exchange of Ca, Mg, and Fe (Huebner, 1980). The second process is the inversion of pigeonite, whereby pigeonite transforms to orthopyroxene + augite (Huebner, 1980; Hess, 1941, 1960). In the Rowland Spring Formation, orthopyroxene (hypersthene) and augite (ferroaugite) coexist in some of the samples. Additionally, hypersthene lamellae exist in a ferroaugite host, both indicating the cooling of a primary, prograde pyroxene. The existence of hypersthene + ferroaugite suggests the occurrence of inverted pigeonite (Hess, 1960, Huebner, 1980). Based on phase equilibria by Lindsley (1980), the top of the pigeonite solvus is around 825 °C along the hedenbergite-ferrosillite join between 11.5 kbars and 15 kbars. This suggests that the pyroxenes in the Rowland Spring Formation equilibrated above the pigeonite solvus. Upon cooling, the pyroxenes inverted to orthopyroxene + augite in addition to undergoing intracrystalline exsolution. Thus, the pyroxene phases equilibria
indicate a minimum temperature of metamorphism of 825 °C.

Mineral assemblages from the Rowland Spring Formation allow the use of garnet-pyroxene exchange equilibria to estimate pressures during prograde metamorphism. Although many samples contain the appropriate mineral assemblage (see Table 1) only one of these samples has been analyzed using the electron microprobe. This sample, 95AD11, contains the equilibrium mineral assemblage of: garnet + orthopyroxene + clinopyroxene + plagioclase + K-feldspar + quartz. Based on the coexistence of garnet and orthopyroxene, pressures are estimated using reaction C and equation 7 and 8 (Newton and Perkins, 1982, eq. 12; Newton, 1983). Pressures ranged from 12.2 kbars to 13.4 kbars with a geometric mean of 12.9 kbars and a standard deviation of 0.6 kbars (Table 6). As previously discussed, errors in ΔG of the reaction are estimated to be ±1.5 kbars; therefore, the pressure of metamorphism for the Rowland Spring Formation may range between 10.7 kbars and 14.9 kbars.

Using the calculated temperatures and pressures for the Rowland Spring Formation, a P-T diagram was constructed to show the ranges in the estimated prograde metamorphic conditions as shown in Figure 8. Based on this diagram, the Rowland Spring Formation is within the stability field for ferrosillite, α−quartz, orthoenstatite, kyanite, and sillimanite. Additionally, β−quartz and fayalite are stable; however, these phases have not yet been found in samples of the Rowland Spring Formation.

Assuming a standard geothermal gradient of about 25 °C/km, the Rowland Spring Formation would have been metamorphosed at a depth of 30 km to 35 km. Indicated pressures are consistent with these depths.
Given the high temperatures and pressures determined for the Rowland Spring Formation, one might expect partial melting to have occurred. Along the Vulcan Materials quarry road, the Rowland Spring Formation is nearly migmatitic; however, in most outcrops there is little evidence of partial melting. The lack of evidence for partial melting suggests that the Rowland Spring Formation was metamorphosed under relatively anhydrous conditions.

In addition to the prograde mineral assemblages shown on Table 1, both chlorite and sericite occur locally. Both chlorite and sericite are generally fine-grained and represent alteration of the primary prograde equilibrium mineral assemblage. The growth of chlorite and sericite can be attributed to 1) retrogradation during the cooling stages of the peak granulite-grade metamorphism; 2) alteration of the primary assemblage by metasomatic fluids streaming off of the porphyritic, water-rich Corbin intrusive; or 3) a second, lower temperature, prograde metamorphic event. Heatherington and others (1996) and Odom and others, (1973) have dated the Corbin granite between 1.1 Ga and 1.5 Ga. Cross-cutting field relationships show that the Rowland Spring Formation predates the Corbin granite. Thus, the granulite-facies metamorphism is older than 1.1 Ga to 1.5 Ga. Textural relationships between the sericite and chlorite suggest that the sericite formed prior to the development of the chlorite. Given this relationship it appears that the sericite may have formed as a post-granulite phase by K-rich metasomatic fluids that were streaming off of the main porphyritic phase of the Corbin granite. Later, the entire Rowland Spring Formation and Corbin granite were metamorphosed during a Paleozoic metamorphic event. The Paleozoic metamorphism appears to have been a much lower temperature and lower pressure event, probably reaching lower- to upper-amphibolite facies as indicated by garnet-rim temperatures. Chlorite alteration of primary metamorphic mineral assemblages is attributed to this Paleozoic metamorphic event.

ACKNOWLEDGMENTS
We gratefully acknowledge Mike Higgins, Ralph Crawford, and John Costello for their field assistance and introducing the primary author to complexities of the Cartersville District. We thank James J. McGee of the U.S. Geological Survey in Reston, Virginia for guiding us with the electron microprobe analyses. Additionally we thank Deana Sneyd with Golder Associates, Inc. in Atlanta, Georgia for a critical review of the manuscript and many helpful comments to improve the readability of the text. Financial support was provided by the University of West Georgia grant SRG 1053-133-101-000-51200, FRG 10-21-1-14-120-000, and FREG 10-21-1-13-119-000; the Department of Geology, University of West Georgia; the U.S. Geological Survey; and Petrologic Solutions, Inc.. All are gratefully acknowledged.

REFERENCES CITED


Effect of grossular-content in garnet on the partitioning of Fe and Mg between garnet and biotite; an empirical investigation on staurolite-zone samples from the Austroalpine Schneeberg Complex: Contributions to Mineralogy and Petrology, v. 92, n. 3, p. 393-399


INTRODUCTION

The Cartersville district is the oldest continuously active mining district in the Southeastern US (Bearden, 1990) with production of gold, graphite, barite, bauxite, iron and manganese oxides, umber and ocher. Related deposits occur elsewhere in Bartow, Polk and Floyd counties and although outside the main mining district, may throw light on their origin. The eastern and southern part of the district is part of the Blue Ridge physiographic province, while the northwestern part belongs to the Valley and Ridge (Figure 1). The division generally coincides with the Great Smoky (Cartersville) fault in the east and the Emerson (Talladega) fault to the south. Although complicated in detail, the boundary separates highly deformed metamorphic rocks in the east and south from sedimentary rocks, or sometimes low-grade metamorphic rocks to the northwest. Gold and graphite are restricted to the higher grade metamorphic rocks, while barite, bauxite, manganese, umber, and ocher occur with the sedimentary rocks. Iron ores are associated with both terrains. This review is mainly concerned with mineralization in the sedimentary rocks rather than with gold and graphite which are directly related to metamorphism.
GENERAL GEOLOGY

Figure 2 illustrates the stratigraphic section in the Cartersville area. Basement rocks including the Corbin Metagranite and Ocoee metasediments (Proterozoic) have been thrust over younger early Paleozoic sedimentary rocks and are therefore out of stratigraphic sequence. Northwest of the Emerson and Great Smoky faults, clastic rocks of the Chilhowee Group rest unconformably on concealed basement rocks. These are overlain in turn by the Shady Dolostone, sandstone and shale of the Rome Formation, shale and carbonates of the Conasauga Group and cherty carbonates of the Knox Group. Kesler (1950) differs from most workers in restricting the name Shady to the basal part of the older carbonate unit and assigning the upper part to the Rome. Here, we follow the more common practice of assigning the full thickness of carbonates to the Shady and drawing the base of the Rome at the base of clastic facies. We concur with Kesler (1950) that the Shady-Rome contact is diachronous. With the exception of Lower Ordovician beds in the upper part of the Knox Group all these strata are Cambrian in age.

THE ROLE OF WEATHERING, HYDROTHERMAL ALTERATION AND PRIMARY DEPOSITION IN FORMATION OF THE ORES

A major problem encountered in the interpretation of the ore bodies in Bartow, Polk and Floyd counties is the high degree of weathering and the thickness of residuum and colluvium, which obscures fresh rock. With the exception of barite, all the ore deposits associated with the sedimentary section are oxides or hydroxides of relatively immobile elements (aluminum, iron and manganese). Ores of goethite-limonite, manganese oxide and bauxite occur in veins and pockets particularly over carbonate rocks. Goethite-limonite occurs in the residuum of both carbonates and adjacent metamorphic rocks. All deposits were mined from residuum or colluvium and both mineralogy and distribution suggest that weathering played a critical role in formation.

In northwest Georgia, mineralization occurs in two main carbonate units, the Shady Dolostone (Cambrian) and Knox Group (Cambro-Ordovician). The occurrence of barite and associated sulfides within massive dolostones of the Shady and Knox formations is similar to Mississippi Valley type (MVT) ore deposits described from the east side of the Appalachian basin from Alabama to Newfoundland and elsewhere (Kesler & van de Pluijm, 1990). Although mineralogy varies from district to district, characteristic ore minerals in this type deposit include galena, sphalerite, barite and/or fluorite associated with pyrite. Typically, these ores are associated with massive dolostones which are believed to have been invaded by basinal brines, perhaps associated with hydrocarbons, originating in the deeper parts of the Appalachian basin (now concealed). These carbonates served as paleoaquifers as a consequence of high porosity and permeability.

![Figure 2 Generalized stratigraphic column for the Cartersville-Cedartown area.](image-url)
created by dolomitization and the presence of paleokarst as well as their chemical reactivity to acidic brines. Collapse breccias are common, either as a result of preexisting karst or because of solution by hydrothermal fluids. The driving mechanism for the expulsion of these brines is thought to be tectonic loading during one or more of the Appalachian orogenies (Kesler and others, 1995, 1996, 1997; Leach and others, 2010).

The basal unit of the Shady Dolostone is particularly mineral-rich with deposits of iron and manganese oxy-hydroxides not only in Georgia but also in Tennessee and Virginia (King & Ferguson, 1960; Force, 1991). These deposits are closely related to an apparent disconformity between the Shady and Chilhowee Group. This together with their widespread distribution suggests a primary sedimentary source of iron and manganese.

Without weathering none of these deposits would have been economic, but what are the relative roles of original synsedimentary mineralization, hydrothermal solutions, and weathering in creation of these deposits?

**MINERALOGY**

**Barite-**

Barium sulfate is the only non-oxide mineral mined in the Cartersville district. It occurs almost exclusively in residuum (and colluvium) from the Shady Dolostone and owes its abundance to its low solubility during weathering. Outcrops of unweathered dolostone are rare but occur as pinnacles exposed at the bottoms of many barite mines. Most significantly these exposed dolostones contain minor small veins and occasionally blocky masses of barite, but the average barite content is only about 2 percent by weight. Generally, barite crystals are white in hand specimen, clear and transparent in thin section with euhedral tabular rosettes (cockscombs) and appear to have grown within voids in the host rock. However, the occurrence of rare barite pseudomorphs after archaeocyathid fossils indicates that minor replacement occurred in the dolostones. Although, associated with sulfides, the majority of barite postdates the sulfides and is relatively free of sulfide inclusions.

Upon weathering, the dolostone is reduced to an ochreous silty clay residuum containing broken masses of barite. Some residuum appears to have been silicified (jasperoid) and contains angular fragments of barite. Weathering concentrates the insoluble barite and allows mining which would be uneconomic in the fresh rock. To produce a ton of barite requires about 4 to 20 cubic yards of residuum.

Although barite is mainly restricted to the Shady Dolostone, minor occurrences as veins have been noted in the Corbin and Ocoee metasediments close to major faults (Kesler, 1950). These have an important bearing on the age of emplacement.

**Sulfides-**

Sulfide minerals have been reported from both the Shady and Knox formations. In the Shady pyrite, sphalerite, galena, chalcopyrite, enargite and tennantite occur in small amounts associated with barite (Kesler, 1950). According to Kesler (1950), the sulfides occur mainly around the margins of the barite near the contact with host dolostone, indicating that sulfides precipitated before barite. Pyrite is also common in some chert and in the deeper parts of many of the brown iron ore mines.

Drilling within the Knox of the Cedartown area indicates breccias associated with Paleozoic karst (Chowns, 1993, 1994 unpublished). Pyrite is especially common as veins and cement but galena, sphalerite, quartz, calcite and dolomite are also present. Mineralization occurs particularly below the unconformity at the top of the Knox Group and within the Chepultepec Formation.

The karst at the top of the Knox Group is related to the unconformity at the base of the Middle Ordovician. Drill holes close to this unconformity in the vicinity of the old iron mines at Oremont, near Cedartown, Polk County reveal Paleozoic caverns filled by windblown sand and sandy shale that drape the limestone floor. Veins and breccias contain sulfides in the fresh rock and resemble rocks from the upper Knox of Tennessee (Hoagland and others, 1965; Misra and others, 1989).

Collapse breccias in the Chepultepec are associated with paleosols and also karst related. Again sulfides are present and the cavernous
cherts and agates so characteristic of residuum from the Chepultepec are interpreted as siliceous flowstones (Chowns, 1994a).

**Iron Oxides**

The principal iron mineral in the Cartersville district is goethite-limonite which occurs in veins and pockets ranging from massive to fibrous and botryoidal, indicating crystallization from groundwater, sometimes in open cavities. It seems to have two different origins, derived in one case from specular hematite and in the other from pyrite.

The basal ~10 m of the Shady Formation, immediately above the contact with the Chilhowee, is especially iron-rich and yields specular hematite in some mines (e.g. Roan Mine). According to Kesler (1950) hematite replaces skeletal debris and sometimes contains relics of ooidal texture reminiscent of sedimentary ironstones. The specular luster indicates low grade metamorphism.

Elsewhere in the Shady where faulting is suspected, pyrite is present at depth leading authors to interpret some of the ores as gossans formed by the weathering of original sulfides (Kesler, 1950: Hurst & Crawford, 1970). Similar gossans are also developed over the sulfides in the collapse breccias in the Knox Group around Cedartown.

Not all the iron deposits coincide with carbonate terrains. Outcrop belts of the Chilhowee and Ocoee groups are also mineralized as well as higher grade metamorphic rocks close to the Great Smoky and Emerson faults. Here goethite-limonite appears to have developed from sulfides deposited in fractured and faulted metamorphic rocks. However once again Tertiary weathering was critical in concentrating and enriching the gossans.

**Manganese**

Manganese dioxide commonly accompanies goethite-limonite and is assumed to be a weathering residue. However, its primary source within the host rock is uncertain. In the Cartersville area it is particularly common in residuum from the Chilhowee and lower part of the Shady Dolostone (McCallie, 1926). Prismatic-tabular pyrolusite and massive cryptomelane have been identified by x-ray diffraction (Kesler, 1950) and occur as flattened, grape-like nodules (up to 0.5m³) with traces of original bedding. Manganese is also often associated with goethite in the Knox Group (Watson, 1904). Manganese is not known in solid solution in pyrite so it seems more likely the original source was within the hematite or perhaps the carbonate minerals of the Shady Dolostone. Analyses show a small percentage of manganese in most carbonates from the Shady (MnO, 0.01-0.26 weight percent; Kesler, 1950). Force (1991) also reports small percentages of manganese (ave. 490 ppm) at various horizons within the Shady and Knox carbonates of Virginia.

**Ocher and Umber**

Ocher and umber are mined principally from the basal iron rich-beds of the Shady Dolostone (Bearden, 1990, 2008). At first ocher was thought to be associated with quartzite in the Chilhowee Group (Watson, 1906), but Kesler (1950) suggested that the “quartzite” was really a peculiar weathered limonitic chert (jasperoid). Color varies from yellow-brown in ocher to purplish-brown in umber depending on the relative percentages of iron and manganese hydroxides (limonite and wad). Grain size is less than 10 µm. Fe₂O₃ ranges from 55-65 percent and MnO₂ from 0.5 percent in ocher to about 7 percent in umber (Fe 38-45 percent; Mn 0.3-4.4 percent). Based on values in unweathered dolostones given by Kesler (1950), this suggests an enrichment in Mn of around 6-80 times in ocher and umber, respectively, similar in magnitude to enrichment in Fe (65 times).

**Silica**

Quartz is common in veins cutting most rock units close to the metamorphic front and is evidently related to deformation. However, carbonate rocks of the Knox and Shady formations are also commonly replaced by chert. Some chert has a vitreous luster and conchoidal fracture and occurs either as early diagenetic nodules or in association with paleosols and karst. In the Knox some agates appear to be siliceous speleothems. In some cases sulfide minerals are associated with this siliceous karst (Hoagland and others, 1965; Hoagland, 1976; Chowns, 1994).
In the Shady and Conasauga there is also a pervasive, diffuse silicification referred to as jasperoid. It has a rough, earthy luster, hackly fracture and ocherus color. This jasperoid occurs mainly in the residuum from carbonate rocks but not within the fresh rock. It is common in Shady residuum but also occasionally invades sandstones of the Chilhowee Group. In many cases it is brecciated or replaces brecciated dolostone with angular fragments of barite, vein quartz and older silicified dolostone. Where voids occur they are lined by minute euhedral quartz crystals. Kesler (1950) concluded that silicification was directly related to barite mineralization but its absence in fresh rock suggests this silicification occurred during weathering (King & Ferguson, 1960; Cressler and others, 1979).

Bauxite-

The first bauxite discovery in America dates to 1887 and was derived from residuum of the Knox Group near Hermitage, Floyd County adjacent to Bartow County (Watson, 1904). Subsequently, similar deposits were opened up south of Rome and also in the Rock Run district of Alabama (Cloud, 1967). All occur as pockets or pipes in the cherty residuum of the Knox dolostones. Some deposits are pisolithic or ooidal, others structureless. Many deposits are bedded and without the cherty debris characteristic of Knox residuum. This lack of normal cherty residuum was at first perplexing (Hayes, 1895; Watson, 1904) but is explained by the development of sinkhole lakes that concentrated the fine-grained suspension load derived by weathering. The role of sinkholes was made clear by drilling in the Rock Run area of Alabama (Cloud, 1967) and is clearly illustrated at the Gray fossil site within the Knox Group of northeast Tennessee (Zobaa et al, 2011).

Fossil plant debris recovered from lignite in the Booger Hollow bauxite mine in Floyd County south of Rome, indicates a probable Paleocene or Eocene age for these deposits (White & Denson, 1966; Cloud, 1967).

**PARAGENESIS**

**Synsedimentary Minerals**-

The host rocks for the majority of orebodies in Bartow, Polk and Floyd counties are massive dolostones formed by the replacement of limestone. Both rock types are highly soluble and it is likely that the development of karst soon after deposition or in association with the unconformity at the base of the Middle Ordovician provided porosity and permeability for the passage of hydrothermal solutions. Carbonates commonly form solid solutions and small amounts of iron and manganese have been detected in both the Shady and Knox formations in Virginia (Force, 1991). Iron was particularly abundant at the base of the Shady, perhaps related to a disconformity. In places specular hematite has been recorded indicating an original ooidal sedimentary rock that underwent low-grade metamorphism (Kesler, 1950).

**Mississippi Valley Type Deposits**-

The presence of barite together with sulfides in the Shady and collapse breccias with sulfides in the Knox suggests they are Mississippi Valley type deposits and that these units acted as conduits for the migration of the connate water, thought to be the source of mineralization (Kesler and others, 1997). Fluid inclusion studies from barite in the Shady indicate emplacement temperatures between 126-297 °C, equivalent to burial depths of about 1.7 km. (Rife, 1971). Elsewhere it is not uncommon for barite to be associated with galena, sphalerite, pyrite and fluorite in Mississippi Valley type deposits. The widespread occurrence of Paleozoic karst in the Knox Group indicates that this may be the source of porosity and permeability in these massive carbonates. It is unknown whether the Shady was similarly karstic, but the occurrence of large masses of void filling barite would support this hypothesis. Alternatively, void space may have been created by the same hydrothermal solutions that precipitated the barite. Unfortunately very few unweathered outcrops survive and little drill core has been examined. Void space may also have been created by faulting and fracturing.

Since the paleokarst in the Knox Group is pre-Middle Ordovician and if the transporting brines
were driven by tectonic loading the date of mineralization lies between Middle Ordovician and Permian. The occurrence of bedding-concordant detrital sphalerite in Early Ordovician karst (Hoagland, 1976) supports a relatively early age; prior to deformation and before occlusion of karst porosity. However, a combination of radiometric and paleomagnetic dating suggests that deposits from the Knox Group in east Tennessee are either middle or late Paleozoic (Acadian or Alleghenian) (Leach and others, 2001). Based on stratigraphic thicknesses in the southern part of the foreland basin two episodes of tectonic loading are identified during the Taconic and Alleghenian orogenies but not during the Acadian.

The ultimate source of mineralizing brines may be from connate water within the carbonate rocks or from adjacent clastic facies. Kessler and others (1988) and Saunders and Savrda (1993) have suggested Middle Ordovician black shale facies (Rockmart Slate) as a possible source of mineralized brines in eastern Tennessee and near Cedartown, Georgia.

Mineralization in the Chilhowee as well as metamorphic rocks east of the Great Smoky and Emerson faults is probably related to faulting and shearing. However, brines may also have been sourced through the Shady paleoaquifer, which underlies the Corbin Metagranite and Ocoee metasediments, in the foot wall of the Great Smoky fault (Kath, this guidebook). In this case hydrothermal solutions may have been newly expelled from the deeper parts of the Appalachian basin or locally remobilized from older deposits. Clearly the mineralization of the faults and fractures along the thrust front is related to Alleghenian events but mineralization in the paleokarst might be older.

Tertiary Weathering-
All the ore bodies in the Cartersville-Cedartown area are located in residuum (or colluvium), especially from massive carbonates. In places, residuum and overlying colluvium may be more than 75m thick and it is evident that the area has been subject to weathering over a long period (Reade and others, 1980). Judging by the high concentrations of iron and especially manganese in small deposits these elements were evidently mobilized and redistributed locally during weathering (Force, 1991).

A number of workers have commented on an apparent elevation control shown by the location of ore bodies. Bauxite and barite deposits are said to cluster between about 250-350 meters above mean sea level, perhaps related to the level of planation during the Tertiary (Watson, 1904). Whether, the kaolinitic clays and bauxites are remnants of more widespread deposits or were always isolated in sink holes is unknown. Plant fossils from lignite deposits associated with bauxite deposits indicate a Paleocene or Eocene age.

OUTSTANDING PROBLEMS

In writing this review we have glossed over many questions regarding the Cartersville mining district. Some of these questions concern the chemistry of the hydrothermal solutions and details of the driving mechanism that are hotly debated and outside the scope of this paper, but some are directly related to the Cartersville area.

Because of a lack of fresh exposures the relative importance of hematite versus pyrite as the source of goethite-limonite is still unclear. While hematite is favored as a source in the basal Shady, pyrite is preferred in the Knox and in faulted Chilhowee. How much iron was present in the original host rock and how much delivered by hydrothermal solutions? Is all the pyrite hydrothermal? Similarly, what is the original source of manganese?

The Mississippi Valley deposit model provides an attractive explanation for the migration of sulfides and barite into the Cartersville district but fails to explain the detailed distribution. Why is barite restricted to certain zones within the Shady while pyrite is most abundant in the underlying Chilhowee? If both were carried by the same hydrothermal solutions, why are they not more intimately associated? Why is most barite free of sulfide contaminants? On the other hand if they arrived separately and at different times, why was one host rock favored over another and how was porosity generated and retained?

Of the brecciation observed in the carbonates and especially the Shady Dolostone how much is related to Paleozoic karst, how much to solution
by hydrothermal action and how much to solution-collapse during weathering?

How many episodes of silicification are involved? We have suggested three but without seeing more fresh rock and examining thin sections it is difficult to know. In particular, more work needs to be done to characterize the jasperoid, prove it is a product of weathering, and distinguish it from older chert, either hydrothermal or early diagenetic.

A major problem remains attempting to establish original mineralogy and texture through the veil of Tertiary weathering. Based on the relative percentages of iron in fresh rock and ocher (0.6 : 40 percent respectively) weathering has reduced the original volume of rock by around 65 times and the resulting residuum has been let down upon a highly irregular karst surface, perhaps tower karst. We would certainly benefit from a suite of core from fresh rock before the last mines close in the Cartersville area.

**CONCLUSIONS**

1) Primary ore minerals in the Cartersville district include hematite, barite, and various sulfides. With the exception of hematite, which is probably of sedimentary origin all these original minerals are hydrothermal, probably Mississippi Valley Type deposits.

2) Massive dolostones in the Shady and Knox formations are the principal host rocks with porosity and permeability probably supplied by karst formed soon after deposition. Evidence of paleokarst is best seen in the Knox Group around Cedartown.

3) Although mineralization is most prevalent in paleoaquifers, it is not restricted to them. Sulfide, especially pyrite is widespread in the Chilhowee Group as well as in metamorphic rocks in the hanging wall of the Great Smoky and Emerson faults. In this case porosity seems to be related to faults and fractures.

4) Based on the Mississippi Valley Type model, the most likely source of barite and sulfides is from brines expelled from the Appalachian basin, and the most likely driving force tectonic loading. Two episodes of tectonic loading stand out; during the Late Ordovician and subsequently when the Great Smoky and Emerson thrust sheets were emplaced in the late Paleozoic. The mineralization associated with faults and fractures along the metamorphic front must be related to the latter, but karst mineralization is probably earlier.

5) Variations in mineralogy between different host formations and districts may be related to variation in source and possibly age. Some early Paleozoic ore minerals located in karst may have been remobilized and migrated into faults and fractures during late Paleozoic deformation.

6) Most of the ores mined in the Cartersville district are weathering residues derived from the primary ores during the Tertiary. This includes goethite gossans derived from the sulfides and residual barite. Some iron and manganese were likely present in sedimentary ironstones or in solid solution within the carbonate host and were concentrated by intense weathering.

7) Three episodes of silicification are recognized. The first is early diagenetic and responsible for the discrete chert nodules in the Shady Dolostone. A second generation of silica is associated with the emplacement of the sulfides and therefore hydrothermal. Finally, ocherous residuum and collapse breccias were replaced by jasperoid generated by weathering.

8) Tertiary weathering in the Knox produced sink holes that were infilled with kaolinitic clay residues that were sometimes converted to bauxite. Where Tertiary karst intersected mineralized Paleozoic karst, limonite and manganese accompanied kaolinite and bauxite.

9) If the residual deposits associated with the Shady around Cartersville are related to the kaolinite-bauxite deposits, which occur over the Knox Group in Bartow, Polk and Floyd counties, they also owe their significance as ore bodies to Paleocene-Eocene weathering.

**REFERENCES CITED**

Bearden, S. D., 1990, Occurrence and exploitation of ore deposits in the Shady Dolomite, Cartersville Mining District, Georgia; p. 9-12, in Proceedings 24th Forum
Chowns, T. M., 1993, Depositional environments and diagenetic history of the upper Knox Group in core holes from near Cedartown, Polk County, Georgia; unpublished report prepared for the Georgia Geologic Survey.
Chowns, T. M., 1994, Stratigraphy of the Knox Group in coreholes from near Cedartown, Polk County, Georgia; unpublished report prepared for the Georgia Geologic Survey.
Hoagland, A. D., 1976, Appalachian zinc-lead deposits; in Wolfe, K. H., ed. Handbook of stratobound and stratiform ore deposits; v. 6, p. 495-534.
McCallie, S. W., 1900, A preliminary report on a part of the iron ores of Georgia (Polk, Bartow and Floyd counties); Georgia Geological Survey Bull. 10A, 190 p.
Sweetwater district, Tennessee; Carbonates and Evaporites, v 4, p.211-230
DAY 1, Saturday, October 10, 2015:

<table>
<thead>
<tr>
<th>MILEAGE</th>
<th>DESCRIPTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0/0.01</td>
<td>Begin at driveway of Clarion of Cartersville and Mineral Museum Dr., turn right onto Mineral Museum Dr. then turn right onto U.S. Highway 411.</td>
</tr>
<tr>
<td>0.15/0.15</td>
<td>Turn left for southbound ramp of I-75 (toward Atlanta).</td>
</tr>
<tr>
<td>3.50/3.35</td>
<td>Exit right on ramp for GA Highway 20, Exit 290.</td>
</tr>
<tr>
<td>3.80/0.30</td>
<td>Turn left (east) at bottom of ramp on GA Highway 20, cross under I-75.</td>
</tr>
<tr>
<td>4.10/0.30</td>
<td>Outcrops of Chilhowee Group, Nichols and Wilson Ridge Formations on right.</td>
</tr>
<tr>
<td>5.45/1.35</td>
<td>Intersection of GA Highway 20 and Rowland Spring Road. Continue east on GA Highway 20.</td>
</tr>
<tr>
<td>5.95/0.50</td>
<td>Intersection of GA Highway 20 and Vulcan Materials Company Road. Turn right (south) on Vulcan Materials Company Road.</td>
</tr>
<tr>
<td>6.15/0.20</td>
<td>Stop buses and unload.</td>
</tr>
</tbody>
</table>

STOP 1: ROWLAND SPRING TYPE LOCALITY

Stop Leaders: Randy L. Kath and Thomas J. Crawford

This is the type locality of the Rowland Spring Formation of the Proterozoic Allatoona Complex. If we were to walk to the top of the hill we would pass through Rowland Spring Formation and cross the contact with the Red Top Mountain Formation, and then cross into the Corbin Metagranite. Because of the dense vegetation along that traverse, we will focus on the well exposed Rowland Spring Formation.

The Rowland Spring Formation at this locality contains two major distinctly different lithologies: one is a coarse-grained phase and the other is a fine-grained phase. The coarse-grained phase is the dominant lithology and is a light-colored, relatively coarse-grained, porphyroblastic-garnet, biotite, quartz, feldspar, granofelsic-gneiss, with minor crystalline graphite and abundant sulfides. The fine-grained phase generally occurs within the coarse-grained phase is a thin, dark-colored, weakly- to strongly-zoned, orthopyroxene, ± clinopyroxene, quartz, ± garnet, feldspar granofels. These dark-colored granofels bodies are lenticular, and generally oriented with their long axes parallel to foliation, which is parallel to compositional layering.

\[1\] Mileage in the small font represents distance from the last stop.
STOP 2: CORBIN METAGRANITE

Stop Leaders: Jim DeCinque, Rodney Matiak, Jessica Scott and Brandon King

Excellent exposures of fresh Corbin Metagranite crop out along Vulcan’s quarry road. At this locality, the Corbin Metagranite lacks any preferred orientation of feldspar megacrysts. The lack of preferred orientation indicates that the Corbin is truly a granite. We suggest that where preferred orientation of feldspar megacrysts in the Corbin Metagranite occurs, it is due to movement along ductile shear zones.

The Corbin Metagranite here is a coarse-grained rock with exceptionally large phenocrysts of white microcline, plagioclase (~An$_{20-35}$), coarse-grained blue quartz, and a fine-grained matrix consisting of biotite, muscovite (sericite), garnet, and chlorite. Banding in the rock is poorly defined, but locally occurs as stringers and segregations of blue quartz and/or biotite. However, the banding is not a through-going feature in the rock. Garnets occur as anhedral to subhedral grains that are highly fractured and concentrated in biotite-rich zones. Along the fractures, chlorite occurs as an alteration product. Chlorite also occurs as an alteration product of biotite. Locally, both microcline and plagioclase are altered to fine-grained white mica.

Accessory minerals include ilmenite, apatite, zircon, and monazite. Generally, the zircons exhibit a rounded core with an euhedral, zoned overgrowth. The rounded zircon cores suggest a metasedimentary origin for the Corbin Metagranite.

Biotite is ubiquitous to all samples examined from this locality. In thin section, biotite has deep brown to yellow to green pleochroism indicating a high Ti-content. Inclusions of rutile, apatite, and ilmenite are common.
LUNCH STOP

STOP 3: OCOEE METASEDIMENTS AND YELLOW BREECHES MEMBER (?) OF THE WILHITE FORMATION; ALLATOONA DAM ABUTMENT

Stop Leaders: Randy L. Kath, Thomas J. Crawford, and Bill Witherspoon

From the pavilion, follow the woods road south along the creek. Good exposures of metaconglomerate, quartzite, schist and phyllite are exposed in the creek bed and along the east bank. Lithologies are characteristic of the Wilhite Formation of the Walden Creek Group of the Ocoee Series and have undergone low grade metamorphism (Gore and Witherspoon, 2013). The Ocoee Series is an upper Precambrian clastic wedge that forms most of the Great Smoky Mountains of eastern Tennessee and western North Carolina. The Wilhite Formation has the greatest variety of rock types of the Ocoee sequence; it is the only formation that contains an appreciable amount of carbonate rock (Hanselman and others, 1974).

The carbonate rocks in the Wilhite Formation are generally in the form small discontinuous bodies completely surrounded by meta-pelitic rocks. These discontinuous carbonate bodies have been interpreted to represent olistoliths in the Yellow Breeches Member of the Wilhite Formation (LePain, 1987). Where the woods road crosses a large culvert, there are good exposures of one of these carbonate bodies in the creek downstream of the culvert. At this outcrop, the carbonate is a dark-gray, weakly metamorphosed, and fine-to medium-grained marble. Higgins and others (1996) interpreted these carbonates as metamorphosed Conasauga Formation that are exposed through a small structural window, just south of the Coopers Furnace Window.

The contact with the surrounding schist and phyllite is extremely sharp which might be indicative of a fault contact. However, if this carbonate is an olistolith, the contact with the surrounding rock would expected to be sharp.

The occurrence of Conasauga Formation would be highly unlikely in this area. The Corbin Massif and Ocoee Series rocks have been thrust over the lower Cambrian (Chilhowee and Shady) part of the foreland fold and thrust belt. This is illustrated on cross sections A-A’ and B-B’-B” on Plate 1. The nearest Conasauga Formation occurs approximately 3.7 kilometers west, in the footwall of the White Fault, see Plate 1.

Approximately 80 pounds of this carbonate was sampled and processed in an attempt to find any non-calcareous fossils. Dr. Johnny Waters at Appalachian State University processed this sample and dissolved the limestone in acetic acid, but no fossils were found. However, Higgins and others (1996) reported the presence of tiny deformed phosphatic skeletal fragments from carbonates in the Coopers Furnace Window. The presence of these skeletal fragments suggested to Higgins and others (1996, p. 22) that the rocks were Paleozoic and they assigned them to the Lower Cambrian Shady Dolomite. In the Smoky Mountains, Unrug and Unrug (1990) found trilobite, ostracod, bryozoan, and microcrinoid fragments in the carbonates of the Wilhite. Based on this fossil assemblage, they assigned Silurian as the oldest age limit for the Walden Creek Group. A Silurian or younger age is highly unlikely for these carbonates, as there are no known Silurian rocks exposed in the Cartersville District, and the entire Allatoona Complex has been thrust over the basal Cambrian stratigraphy.

From the carbonate exposures in the creek, walk downstream toward the pavilion. Just north of the pavilion there are good exposures of metaconglomerate, quartzite, and schist along the west side of the road. At the intersection with Allatoona Dam Road, Higgins and others (1996) mapped a small
lens of Shady Dolomite. Although we have not been able to substantiate the occurrence of a carbonate at this intersection, if it does occur here we would interpret this as another small lens of carbonate within the Wilhite Formation.

Turn east on Allatoona Dam Road, there are very good exposures of the Wilhite on the south side of the road. At the fence to the Allatoona Dam Power House, there are good exposures of metaconglomerate overlying phyllite/schist that are underlain by metagraywacke. This exposure was photographed and described by Gore and Witherspoon (2013, p. 198). Bedding is well preserved in the meta-sediments. The bedding is undulatory and dipping to the northeast. The schist/phyllite layer between the metaconglomerate and metagraywacke show some low-amplitude folds that are characteristic of flexural shear between the more competent layers. There are well developed crenulations in the phyllite/schist layers and a well-developed cleavage. The bedding-cleavage angle suggests that portions of the flexural slip folds are overturned to the north-northeast.

Unfortunately, for this fieldtrip we were not allowed access inside the secure area around the dam. If we had access to the stairs along the left (south) abutment of the dam, we would have traversed up the stairs to the top of the dam. Many of the rocks in the section adjacent to the stairs have been described as mylonites (Higgins and others, 1996). About half way up the stairs, there is a westward-dipping block of mylonitized Corbin Metagranite. The Corbin Metagranite is followed, going up stairs, by quartzite and metaconglomerate of the Wilhite Formation. The Wilhite is followed by another block of Corbin Metagranite. On top of the Corbin is a gray phyllonite mylonite with pods and lenses of quartz, and with pods of what appear to be intensely mylonitized Corbin.

Higgins and others (1996) described the meta-sedimentary rocks of this area as part of the Chilhowee Group and Pinelog Formation. They interpreted the isolated blocks of Corbin as part of a duplex. Furthermore, they interpreted the shearing in the Corbin to be associated with an eastward dipping thrust fault that separates the Corbin Massif from the Chilhowee Group and Pinelog Formation. This fault has been named the Allatoona Dam Fault.

Based on detailed geologic mapping under the Allatoona Dam power house and left (south) abutment by Crawford and others (2009), the Allatoona Dam Fault is a high-angle, west-dipping normal fault that separates the Ocoee Series rocks from the Corbin Massif. Based on age relationships and geometry of the fault, we currently interpret this fault to be related to a Rodinian rift basin (graben) which preserves Ocoee Series sediments.

Retrace route back to buses.

20.9/0.00 Board buses. Retrace route west along Allatoona Dam Road to U.S. 41.
22.4/1.50 Intersection of Allatoona Dam Road with U.S. 41. Turn left (south) on U.S. 41.
23.6/1.20 Turn left (southeast) onto Red Top Mountain Road connector
23.8/0.20 Turn left (northeast) onto Red Top Mountain Road Southeast
25.4/1.60 Cross Bethany Bridge, continue on Red Top Mountain Road
27.3/1.90 Turn into the Iron Hill Trail parking lot. Unload busses.
STOP 4: GEOLOGY OF THE IRON HILL CAMPGROUND AND WALKING TRAIL AT ALLATOONA LAKE, BARTOW COUNTY, GEORGIA: STRATIGRAPHIC AND KINEMATIC EVIDENCE FOR SEPARATION OF THE CARTERSVILLE-GREAT SMOKY AND EMERSON-TALLADEGA FAULTS

Stop Leaders: Randy L. Kath and Thomas J. Crawford

**Introduction**

We will park the busses in the new parking lot for the Iron Hill trail system which is part of Red Top Mountain State Park. After unloading the busses, please gather at the trailhead near the trail map sign at the west end of the parking lot. Once at the trail head sign, we will start our traverse to the abandoned boat ramp at the Iron Hill Campground. Please try to stay together during this traverse.

The general location of the walking traverse is shown on Figure 1. The green line represents the traverse from the busses to the cove and the red line represents the return traverse. There are several notable areas that are flagged with solid blue survey flagging along the trail. Each of these notable areas is described in the following text. We will eventually end up at the Crowe Bank and Allatoona Bank iron mines. These two mines have been consolidated and are now called the Iron Hill Mine (McCallie 1900; Haseltine 1924; Hurst and Crawford, 1970; Chowns and Kath, 2015 (*this volume*)).

The Iron Hill Mine combines two contiguous deposits that follow the strike of the Chilhowee Group rocks that underlie Iron Hill, the Crowe and Allatoona Banks, separated by a narrow barren interval where the ore pinches out. Ore occurs as weathered veins and breccia infill in phyllite and quartzite of the Chilhowee Group close to the contact with the Corbin Metagranite (Hurst and Crawford, 1970; Crawford and others, 2009; Crawford and others, 2010; Kath and Crawford, 2011). The material appears to have originated as pyritic fault breccia, which was subsequently oxidized to a limonitic gossan (Kesler, 1950). Without the effects of deep weathering the deposit would not have been economic.

The Allatoona Ore Bank was originally opened before the civil war by Mark A Cooper to supply ore for a small furnace on Allatoona Creek (McCallie, 1900). “After the war the bank was operated for some time by an Atlanta Company. This company constructed a narrow gauge branch railroad, a mile and three quarters long, connecting the ore bank with the Western & Atlantic R.R near Allatoona Station. The ore-bank finally passed into the hands of the Etowah Iron Co. which operated it until [1902]… The main workings at the Allatoona bank consist of two open-cuts near the top of the ridge. These excavations cover a total area of about three quarters of an acre. They are each about 100 yards long, and from 10 to 60 feet in depth. The cuts are entered by tunnels driven on a level with the railroad, which runs along the side of the hill. By this means the ore is conveyed directly to the cars, without having to be hoisted from the cuts. The orebody … seems to partake of the nature of both vein and pocket deposit”.

By 1910 the local blast furnaces were closed and ore was being shipped to South Pittsburg and other localities in Tennessee (Haseltine, 1924). The Crowe Bank continued to be worked until 1923 (Kesler, 1950). By this time, the excavation was about 130 feet deep and equipment consisted “…of a 4-log washer, three steam shovels, three dinkey engines, and a small locomotive…. At the time… 40 men were employed and about 4 cars of ore per day were being shipped. Some of the ore went to Rockwood, and LaFollette, Tenn., and the rest to Gadsden, Alabama” (Haseltine, 1924). Most of the mines were closed by this time probably due to a lack of coking-coal, the closing of local blast furnaces and the expense of transport. A couple of mines continued to operate until the end of World War II (Kesler, 1950).
Walking Traverse-

The parking lot area is completely underlain by the Corbin Metagranite of the Allatoona Complex (Crawford and others, 2009; Kath and others, 2015 (Plate 1 this volume)). The exposed Corbin in the parking lot is highly sheared and contains a well-developed shear fabric with porphyroclastic feldspars with sigmoidal tails. The shear fabric generally trends 020/64 (N20E 64SE), which is the general orientation of shear fabrics developed in the Allatoona Complex east of the Allatoona Dam Fault and north of the Emerson-Talladega Fault (see Plate 1). Corbin residual soils are generally comprised of white to yellowish, clayey-sand, and contain coarse-granular quartz. Much of the quartz is blue and has a waxy to greasy appearance. The blue color in quartz has generally been attributed to either Ti$^{4+}$ substitution for Si$^{4+}$ or as the result from Rayleigh scattering caused by tiny inclusions (Wise, 1981). When we return to the busses after walking the traverse, we will inspect several outcrops of the Corbin, if time permits.

As we traverse along the trail, note the exposures of light colored soils that are characteristic of the Corbin. Upon close inspection of the soils, very coarse-grained quartz fragments can be observed, many of which have the characteristic blue color. At the trail/road intersection, turn south-southwest (right) and follow the old paved Iron Hill Campground road. This road eventually ends at the old boat ramp, where we will start our lake shore traverse.

As we continue along the old road bed, the first exposures in the road cuts are of the sheared Corbin Metagranite. At the first blue flag, along the east side of the road, there is a small exposure of the sheared Corbin. Many of the fragments appear to be “phylite” but are actually phyllonite derived from the intense shearing of the Corbin. Look closely with a hand lens and you will see the characteristic coarse-grained, blue quartz in the phyllonite. Larger fragments of sheared Corbin, located adjacent to the phyllonite, contain the phenocryastic feldspar, now porphyroclastic, and coarse-grained blue quartz.

After examining this small exposure of sheared Corbin, continue along the road to the south-southwest. A large, angular boulder of Chilhowee Group silicified sandstone (quartzite) is located approximately 100 feet south of the sheared Corbin outcrop. Based on the stratigraphy of the Chilhowee presented by Mack (1980), this would either be equivalent to Wilson Ridge or Weisner Formation. Regardless, this block is not in place and has moved down slope from the ridge top immediately east of the road. The Crowe Bank and Allatoona Bank iron mines (Iron Hill Mine) are just over the top of the ridge east and up slope of this silicified sandstone boulder.

Continue toward the lake on the old road bed. The next stop on this walking tour is along the northwest (right hand) side of the road, near another blue flag. At the flag, one can see mine dumps presumably from the Crowe Bank deposit. In the distance, beyond the mine dumps, the shoreline of Allatoona Lake is very light colored and contains abundant blue quartz. The saprolite exposed on this shoreline is developed from the Corbin. Across the road from this flag and upslope from the road, there are several old mine pits and small dump piles related to the Crowe Bank mine.

At the next blue flag the opening to the Crowe Bank mine is at road level. The outcrops exposed on the far (east) side of the mine contain phyllite and feldspathic sandstone of the Chilhowee Group. Based on mapping throughout the Cartersville District, we interpret these phyllites and interlayered sandstones to be weakly metamorphosed Nichols Formation, using the stratigraphy defined by Mack (1980). Within the Nichols Formation there are abundant layers/beds of siltstone and fine-grained sandstone. Previous workers in the district have collectively called this Weisner Formation. The Weisner was previously defined as all clastic units below the Shady Dolomite and above the Ocoee Supergroup (Wilhite Formation). However, Mack (1980) defined the Weisner Formation as the uppermost sandstone of the Chilhowee Group.
Figure 1. Areal overview of walking traverse to and from the Iron Hill Campground and boat ramp.
Continue downslope to the next blue flag, where there is an old overlook into the Crowe Bank mine. Please use caution when approaching the mine pit; the old platform is very unstable. We suggest walking adjacent to the platform to observe the old mine workings. Excellent exposures of the silicified and mineralized Chilhowee Group rocks occur directly beneath your feet within the highwall. We recommend that you do not try to sample the ore at this locality; there will be plenty of areas along this traverse where better samples can be observed.

Return to the old road bed and continue downslope to a trail intersection sign. At the trail intersection, take the left fork; stay on the old paved road. Shortly after the fork, another trail will turn south toward the lake. Stay on the old paved road and continue to the boat ramp. The Chilhowee/Shady contact occurs just beyond the second trail intersection. As we cross into the Shady Dolomite, several small mine prospects (pits and trenches) can be observed. In these prospects, the soil has the characteristic red and ochre color of the Shady residuum. Many of the prospects contain large boulders of jasperoid, which are silicified Shady carbonate.

There is an abandoned boat ramp at the end of the paved road. At the boat ramp, go down to the lake shore and traverse the shoreline to the southwest. If you are uncomfortable walking along the shoreline, there is a trail at the top of the boat ramp that goes through a series of prospects (trenches and pits) that were excavated in the Shady residuum. If you follow this trail, you will end up in a small cove of Allatoona Lake where the Chilhowee and Corbin are well exposed along the lake shore.

![Figure 2. Areal distribution of rock units exposed in small cove along walking traverse.](image-url)
Along the shoreline southwest of the boat ramp, the Shady Dolomite, Chilhowee Group, and Corbin Metagranite are well exposed (Figure 2). At the bottom of the boat ramp, notice the dark red and reddish brown color of the soils. These soils are characteristic of residuum developed on the Shady Dolomite. Many of the white colored cobbles and gravel fragments that are scattered around the high water level are composed of either vein quartz derived from Blue Ridge rocks or barite weathering out of the Shady Dolomite. As we continue our shoreline traverse, the concentration of barite increases and reaches a maximum concentration near the Chilhowee/Shady contact.

Economic deposits of barite in the Cartersville Mining District occur only in the Shady Dolomite. No other unit within this mining district is known to have economic concentrations of barite. Several well-known localities for barite crystals have been exploited by mineral collectors, but these represent remobilization of the barite and open space crystallization in jasperoid after Shady, or in silicified sandstone/siltstone of the Chilhowee Group.

At the point along the shoreline where the barite concentration is at its greatest, note the general shape of the barite. Most of the barite is somewhat flat and angular; however, a few cobbles and pebbles of barite are sub-rounded and appear to have been transported or reworked by the lake. Above this lake shore exposure, there is a blue flag where examples of both morphologies of barite have been placed on an uprooted tree (Figure 3).

At lake-shore level, the high bank above this barite concentration exposes weathered Shady with angular to sub-angular barite. This exposure is characteristic of eluvium and provides insight into the concentration mechanism of the barite that allowed development of a viable economic deposit. The scarcity of rounded barite indicates that much of the barite is concentrated by in-situ weathering of the Shady, along with gravitational movement and concentration, forming an eluvial stratigraphic horizon (Figure 4). Barite is not known to have been mined from this locality (Bearden, 2015, per. comm.).

Continue walking along the lake shore into a small nearly north-south trending cove. If you look across the lake to the southeast, the shoreline exposes rocks of the Blue Ridge and, most importantly, the Corbin Metagranite. As we round the bend into the cove, the residuum will become much lighter and good exposures of phyllite, graphitic phyllite, and quartzite can be observed. These exposures show characteristic lithologies of the Chilhowee Group. The saprolite and residual soils all contain small “chips” of phyllite and graphitic phyllite, very different than the dark red soils derived from the Shady Dolomite, especially the eluvial horizon.

Kinematic indicators are well preserved in the more micaceous layers of the Chilhowee. The phyllites and graphitic phyllites are highly sheared and have developed strong linear fabrics. Most of the linear fabrics are related to crenulations, but many are slip lineations. Systematic measurement of the crenulation and slip lineations was conducted to determine the average orientation of these linear
features and to evaluate whether they are consistent with strike-slip or dip-slip or oblique-slip motion along any of the mapped faults.

Twenty four lineations were measure in this cove in phyllites and quartzites of the Chilhowee. The average orientation of the lineations is N76W-S76E. Plunges are moderately inclined to the southeast, with an average of 33 degrees (33/104 or 33/S67E).

Continue traversing into the cove and notice the shoreline sediments that are derived from the Chilhowee phyllites. The Corbin Metagranite is exposed at the next blue flag. Note the difference in outcrop character and the grain size increase combined with the presence of blue quartz. Exposures of the Corbin are limited on this side of the cove; however, an exposed knob of Corbin occurs directly north of the blue flag. If the lake level is low enough, we will traverse over to this exposure.

The feldspars in the Corbin are highly sheared and porphyroclastic at this exposure. The tail morphology of the porphyroclasts indicate oblique-slip movement, with a major component of dip-slip movement. Measurement of 14 kinematic indicators in the Corbin, show an average slip direction of N56W-S56E (Figure 5). Plunges are moderately inclined to the southeast, with an average of 39 degrees (39/124 or 39/S56E). The orientation and magnitude of the lineations are consistent with those measured in the Chilhowee. The average orientation of all lineations measured in the Chilhowee and Corbin is N68W-S68E, with an average plunge of 36 degrees (36/112 or 36/S68E).

The lake shore southeast of this cove is also underlain by Corbin Metagranite. Thus the stratigraphy along a northwest to southeast cross section would be Corbin-Chilhowee-Shady-Corbin. The areal distribution of this tectonostratigraphic package is shown by Crawford and others (2009, 2010) and on Plate 1 (this guidebook). We interpret the Chilhowee and Shady contact (see Figure 2

---

---
and Plate 1) to be a normal stratigraphic contact. Because these units are completely enclosed by Corbin, the Shady-Corbin and Chilhowee-Corbin contacts must be faults.

Throughout the Cartersville District and Allatoona Dam 7.5-minute quadrangle, the Corbin Massif and Allatoona Complex are thrust over the folded foreland lower Cambrian stratigraphy, Chilhowee Group and Shady Dolomite, by the Cartersville-Great Smoky Fault. This explains the juxtaposition of the Corbin Massif structurally overlying the basal Cambrian stratigraphy, and explains the Shady-Corbin contact southeast of the cove (Figure 6, B"-B'-B panel).

In this cove, the Chilhowee structurally overlies the Corbin. Although this is a correct stratigraphic sequence (King, 1964; Hurst and Schlee, 1962; Newman and Nelson, 1965; Mack, 1980), the exposed Chilhowee is only about 250 feet in outcrop width; much thinner than anywhere in the Cartersville district. The tectonic thinning and kinematic indicators indicate that this contact is a fault and not an unconformity. This fault is younger than the Cartersville-Great Smoky Fault and brings the “flat” portion of the Cartersville-Great Smoky Fault up to the surface along a “ramp” of the Emerson-Talladega Fault (Figure 6, Panel C-C’)

After examining the Corbin, traverse east through the woods, to the Iron Hill hiking trail. Follow this trail back toward the boat ramp. Good soil exposures of the Corbin occur where we first enter the trail. At the low point of the trail, the soils become redder and contain phyllite fragments of the Chilhowee. Continue up the hill. At the next blue flag, in the sharp turn of the trail, the phyllite of the Chilhowee is well exposed. The contact with the Shady is also exposed along the woods line, immediately south of the phyllites (toward the boat ramp). Notice that there are abundant prospects in the Shady residuum. This part of the Shady was prospected for iron ore.

Continue walking along the Iron Hill trail back to the trail junction. Turn hard right along the old paved road that we have already traversed. At the next trial junction, turn left and walk parallel to the lake shore toward the east.
For those of you that are adventurous, take the first trail down to the lake shore. Once at the lake shore, traverse back to the west along the shoreline (toward the cove that we just left) where there are good exposures of the iron ore. There are several large stock piles of the ore that have been exposed by wave action of the lake (Figure 7). We suspect that this ore was being extracted from the Allatoona Mine, most likely the Crowe Bank. Given the nature of the ore in these stock piles, the ore was probably too silica-rich to be economically viable. Regardless, one can see the nature of the deposits. Several varieties of ore can be seen in the stock piles, including botryoidal, massive, bedded, and breccia infilling.

For those who choose not to see the ore piles, continue along the Iron Hill trail. Along the north side of the trial, you will see another blue flag with several boulders beneath. These boulders are sandstones of the Chilhowee that have been silicified. The original bedding is well preserved and accentuated by differential weathering. Because the Shady and Chilhowee are both silicified, we have made a distinction between silicified carbonate and silicified sandstone and siltstone. The silicified carbonate has been called jasperoid in the mining district. However, many workers (Kesler, 1950) also called the silicified sandstone/siltstone jasperoid. Based on our work, we would restrict the usage of jasperoid to represent silicified carbonate.

After leaving the soil cut, we will continue east along the Iron Hill trail. The entrance to the Allatoona Bank mine is flagged. Please enter the old mine working, but be careful when approaching the high walls.

The first high wall exposure is phyllite of the Chilhowee Group. As you continue into the old mine workings, silicified sandstone and siltstone as well as mineralized fault breccia are well exposed (Figure 8). Locally, wavelite had been found in the Allatoona Bank mine. The wavelite forms botryoidal masses that are yellowish-white in color, as illustrated on Figure 9. The sample shown on Figure 9 was collected nearly 40 years ago by Tom Crawford while mapping the Allatoona Dam
quadrangle (Cressler and Crawford, 1976). An interesting note is that similar wavelite samples, both in form and color, have been found in mine pits associated with Chilhowee fault breccia in the Rock Run/Oremont mining district of eastern Alabama.

Exit the Allatoona Bank Mine, continue along the trail toward the busses. The last exposure along the trail consists of feldspathic sandstone overlain by gray phyllite of the Chilhowee. After leaving this exposure, pay attention to the first occurrence of spherical boulders in the woods. These joint-controlled exfoliation boulders are derived from the Corbin; you have just crossed the Emerson-Talladega Fault.

At the top of the hill, continue across the old paved road and return to the parking lot and load the busses.

References Cited
Crawford, T.J., Kath, R.L., and Costello, J.O., 2009, Bedrock Geologic Map of the Allatoona Dam, 7.5-Minute Quadrangle, Georgia: Georgia Department of Natural Resources Open File Report 09-1, Plate 2
Haseltine, R.J., 1924, Iron Ore Deposits of Georgia: Georgia Geologic Survey Bulletin 41, pp. 238
27.3/0.00  Board buses and retrace route to Interstate 75
30.2/2.90  Merge right onto I-75 north towards Cartersville.
31.5/8.10  Take Exit 293 for US-411 toward Chatsworth/White
38.3/0.30  Turn left (south) onto US-411 South towards Rome
38.6/0.40  Turn left (east) on Mineral Museum Drive.
38.7/0.10  Turn left into the Clarion Inn parking lot

END OF DAY 1.
49TH ANNUAL FIELD TRIP: ROAD LOG

DAY 2, Sunday, October 11, 2015:

<table>
<thead>
<tr>
<th>MILEAGE</th>
<th>DESCRIPTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0/0.0²</td>
<td>Begin at driveway of Clarion of Cartersville and Mineral Museum Dr., turn right onto Mineral Museum Dr. then turn right onto U.S. Highway 411.</td>
</tr>
<tr>
<td>0.10/0.10</td>
<td>Turn left for southbound ramp of I-75 (toward Atlanta).</td>
</tr>
<tr>
<td>8.4/8.30</td>
<td>Exit right on ramp for Red Top Mountain Road, Exit 285.</td>
</tr>
<tr>
<td>8.7/0.30</td>
<td>Turn right (west) at bottom of ramp on Red Top Mountain Road SE.</td>
</tr>
<tr>
<td>10.3/1.6</td>
<td>Turn left to Georgia 293/Emerson</td>
</tr>
<tr>
<td>10.6/0.30</td>
<td>Turn left (north) on Georgia 293</td>
</tr>
<tr>
<td>5.95/0.30</td>
<td>Turn left (west) into New Riverside Ochre’s entrance road</td>
</tr>
<tr>
<td>6.05/0.10</td>
<td>Stop buses and unload at the processing Plant.</td>
</tr>
</tbody>
</table>

STOP 1: NEW RIVERSIDE OCHRE- EMERSON BARITE MINE

Stop Leaders: Stanley Bearden, Randy Kath and Tom Crawford

Introduction
Production of barite from the Cartersville Mining District was first reported in 1894. New Riverside Ochre Company, Inc. (NRO) has been open-pit-mining barite in the Cartersville District since 1924. Barite deposits in the district occur stratigraphically above the ochre deposits, which occur in the lower 10-meters of the Shady. Throughout most of the district, the barite occurs as an epigenetic, stratabound and strataform, structurally deformed, Mississippi Valley Type (MVT) deposit hosted in the Shady. In this model, barium enriched low-temperature hydrothermal fluids migrated along stratigraphic and structural conduits and precipitated in open space in the Shady. Alteration is relatively minor and inconspicuous, with only minor barite replacement of the original carbonate. Most of the ore was deposited in solution cavities, solution enlarged joints, and locally, in fault breccia.

Although the MVT model is widely accepted throughout most of the district, exposures at this property (referred to throughout the text as the Emerson Mine) indicate a secondary concentration mechanism of barite and associated materials. The origin of this deposit is discussed below in the “The Emerson Mine” section.

Barite is grouped into seven general categories based on use: (1) hydrocarbon drilling fluids; (2) extenders and fillers; (3) glass and fiberglass; (4) paint; (5) chemical products; (6) high-density applications, and (7) frictional materials. All barite that is currently mined and processed by NRO is

² Mileage in the small font represents distance from the last stop.
used in the extender/filler, chemical products, high-density markets, and frictional materials. Originally, barite mined in the Cartersville District was used for the manufacture of white paint pigment (lithopone), a mixture of barium sulfate and zinc sulfate.

Principal applications of barite in the filler/extender markets include paint, latex, and urethane foam manufacturing. In painting automobiles, barite is used in the primer coats to retard rusting. It also contributes to the gloss of the top coat. Processors of polyurethane foam use barite in manufacturing such products as floor mats and carpet backings and tennis balls to increase density and improve processing qualities. Other important applications include mold-release compounds in metal casting, brake systems and other frictional materials, acoustical compounds, and in high-density concrete.

Property History
The first barite prospected on this property was conducted in the late 1880’s by Mr. W.R. Satterfield. After the prospecting was completed, barite mining activity was conducted by Thompson-Weinman, The Nulsen Corporation, and E.I. DuPont de Nemours & Company. The initial mining activity on this property ended in 1944, but was reactivated in 1976 when Thompson-Weinman and NRO jointly explored the property which resulted in the identification of three distinct mineral deposits: a northern deposit, central deposit, and southern deposit.

Additional exploration using wet rotary drill holes was conducted between July 1978 and July 1979. Based on this exploration, the northern and southern barite deposits were estimated to be comparable in size and tonnage, and could be mined using open pit methods. Unfortunately, the central deposit is too deeply buried to be economically mined using open pit methods.

Eventually, Thompson-Weinman and NRO divided the property with Thompson-Weinman taking the northern deposit and NRO taking the southern and central deposits. NRO mined the southern deposit between February 1996 and May 2004.

In July 2012, NRO purchased the northern deposit. A mine permit was issued in August 2012, and mining began in January 2013. Based on initial ore reserve estimates, this mine will be the 4th largest barite producer in the entire Cartersville District.

Ore Extraction
Barite is extracted from the Emerson Mine using wheel tractor elevating scrapers and track-type excavators that support the main excavation using a Manitowoc 4600 Vicon dragline (see guidebook cover). The dragline has a boom length of 140 feet, a 4 cubic yard bucket, and has a maximum excavation depth of 90 feet. Rubber tire front-end loaders with 7 cubic yard buckets load 35-ton trucks which transport crude ore matrix to stockpiles or to the washer plant. An average of three acres per year is excavated; removing approximately 500,000 cubic yards of earth per year; about half of which is crude ore matrix.

Barite Beneficiation
Crude ore and associated gangue material are processed by NRO using three different plants:

- Washer Plant
- Floatation Plant
- Magnetic Separator Plant

During this fieldtrip we will visit the washer plant; therefore, the washer plant is described below.
Ore Processing (Washer Plant)
Thirty-five ton trucks deliver payload to the washer plant at approximate 15-minute cycles. Cycle time is dictated by washing characteristics of the crude ore matrix (clay to rock content, contained ore percentage, and nature of gangue).

The function of the washer plant is to produce a slurry from which the following products are generated:

1. Jig ore
2. Scavenger jig ore
3. Hutch/screen sands
4. Akin classifier sands
5. Dornicks
6. Gravel
7. Akin classifier overflow

The washer plant consists of four log washers in closed circuit with screens and crushers. This circuit delivers wet classified sands and gravels to four sets of two-cell 42-inch Bendalari jigs which feed two sets of two-cell 36-inch scavenger Bendalari jigs.

Jig ore is the first identifiable and marketable product derived by the washer plant. This ore ranges between 94% and 97% weight percent BaSO₄. Particle size is from 3/16-inch to 1-inch. Washer plant feed rate and volume are determined by visual assessment of jig ore quality. Scavenger jig ore is the final concentrate from the last cell of the jig sets; it is too highly contaminated with iron, manganese, and silica to be blended with the jig ore and contains too great a weight percentage of barite to be blended with gravel.

In the past, the scavenger jig ore was sold as a drill-mud grade product. Since 1976, scavenger jig ore is recycled onto the crude ore matrix stockpile. BaSO₄ content of this ore will range from 40 to 70 weight percent. Hutch sand is a barite-enriched (30% ± 10%) minus 3/16-inch jig product. Minus 3/16-inch screen sands from up-log discharge are mixed with hutch sand. An Akin classifier removes the plus 50-mesh fraction for the log overflow stream. These sands average 10% weight percent BaSO₄ and are minus 3/16-inch by plus 50-mesh in particle size.

Hutch and Akin classifier sands are truck-transported to the Floatation Plant ball mill stockpile. Dornicks are plus 5 ½-inch boulders rejected by the bull screen prior to log washer feed. This material is stockpiled for future recycle. The oversize material is a minus 5 ½-inch by plus 3-inch steam rejected by the vibrating screen which segregates non-overflow log product. This material is belt-conveyed and can be hand selected upon demand for immediate barite recovery. Rejected material is stockpiled for recycle or as a source of rip-rap. Gravel is a final product from the jigs. Particle size of the gravel from 3/16-inch to 1-inch and barite content 4% ± 2%. Akin classifier overflow is approximately 3,500 gallons per minute, 15% solids slurry which is transported to a hydrocyclone for floatation recovery or to impoundment as final tailings. This stream contains an average of 2 tons per hour of plus 10 micron recoverable barite.

The Emerson Mine
Although a MVT model is called on for most of the barite deposits in the Cartersville District, the barite-concentration processes of the Emerson Mine are unique when compared to other mines in the district. This mine is characterized by thick accumulations of colluvium that are derived from the Chilhowee and Shady. The colluvium locally contains well-rounded alluvial cobbles of vein-quartz that are derived from the adjacent Blue Ridge metamorphic rocks.

The Emerson Mine is one of the southernmost barite deposit in the district. Also, it is one of the mines closest to the Blue Ridge metamorphic front and lies west of the Cartersville-Great Smoky Fault, east of the Cloverleaf Fault, and north of the Emerson-Talladega Fault. Detailed geologic
mapping by Kath and others (2009, 2010) of the Cartersville 7.5-minute quadrangle places the mine on the northeastern limb of a doubly plunging, west-northwest verging anticline. Bedding measured in Chilhowee quartzite dips steeply to the east. This steep dip is responsible for the high topographic relief of the mine site. The topographically highest part of the mine property, west of the open pit, is underlain by quartzite and phyllite of the Chilhowee Group. The ground surface above the mine forms a northeast facing dip slope on the Chilhowee with an average slope angle between 45 and 55 degrees. The slope angle flattens to less than 40 degrees when underlain by the Shady Dolomite.

Barite is mined at the Emerson Mine by open pit methods. The ore zones within the open pit are developed in colluvium that is derived from the upper, light-gray, Shady Dolomite. Pinnacles of the upper, light-gray Shady are exposed in mine pit (Figure 1). These pinnacles are completely surrounded by thick accumulations of colluvium that contain minor alluvial gravel. The unoxidized matrix of the colluvium is medium- to dark-brown, silty clay that contains angular fragments (sand-to boulder-size) of barite, sub-angular to angular fragments (sand- to boulder-sized) of quartzite, and well-rounded vein quartz cobbles and gravel (Figure 2). Locally within the dark-brown matrix there are zones that contain dark-red oxidized clay with similar barite, quartzite, and vein quartz material.

The dark-brown silty-clay material is best exposed near the bottom of the open pit. In the upper parts of the highwall, this dark-brown material has been locally oxidized during recent weathering, producing a characteristic ocher color typical of Shady residuum. This ocher-colored material is considered to be the more deeply weathered equivalent of the underlying dark-brown material based on the similar presence of abundant barite, quartzite, and rounded vein quartz.
Immediately adjacent to most of the Shady paleokarstic pinnacles there is a thin rind of light-brown to ochre-colored saprolitic residuum that contains a fabric sub-parallel to the fresh dolostone. This rind ranges in thickness from 0.1 to 1 meter and does not contain any ore-grade barite, quartzite, or vein quartz. This material is characteristic of Shady residuum developed from in-situ weathering of the light-gray dolostone seen in other mine pits throughout the District.

Overlying the Shady, the upper benches of the open pit expose at least three separate layers of colluvium derived from the Chilhowee. The uppermost, relatively youngest colluvial layer is characterized by a dark-red, silty and sandy clay matrix with abundant angular quartzite fragments. The contact between this relatively younger colluvium and the underlying, relatively older colluvial (middle) layer dips around 32 degrees to the east-northeast, as shown on the left side of the upper bench in Figure 3. The middle colluvial layer is similar to the upper colluvial layer, except that it has far less angular quartzite fragments and is richer in clay matrix. The lowest (oldest) Chilhowee-sourced colluvial layer is exposed above the lower bench shown in Figure 3. This lighter-colored material is characteristic of the light-colored Chilhowee colluvium seen throughout the district.

As stated previously, the Emerson Mine is not characteristic of other barite mines in the Cartersville District. Other barite mines in the district are formed mostly by in-situ weathering of the upper light-gray Shady dolostone. The barite occurs as irregular and sub-rounded masses completely within a light- to dark-brown Shady residuum. At the Emerson Mine, the barite is angular and is completely contained within colluvium. In-situ weathering of the upper light-gray Shady dolostone, combined with down slope movement and gravity...
accumulation of the barite is responsible for this eluvial-style deposit. (Eluvial deposits consist of soils that are derived by in-situ weathering combined with gravitational movement or accumulation of soils.)

Deposition of the eluvium occurred between the paleokarstic pinnacles of the upper light-gray dolostone. During deposition, this area may have resembled a tower karst topography formed on the light-gray dolostone. Because of the proximity to metamorphic rocks, this area was protected from Paleocene and Eocene weathering; however, by late Miocene and early Pliocene, this area would have been deeply weathered. This late Miocene and early Pliocene weathering has been well documented in other ore districts throughout the world. Further, Miocene epeirogenic uplift of the southern Appalachians (Gallen, Wegmann, and Bohnenstieh, 2013) may have caused increased topographic relief and accentuated colluvium development adjacent to the more resistant ridges that were held up by quartzite of the Chilhowee. The topographic setting combined with proximity to the metamorphic front accounts for the unique nature of this barite deposit.

References Cited:
Kath, R.L., Bearden, S.L., Costello, J.O., and Crawford, T.J., 2009, Bedrock Geologic Map of the Cartersville, 7.5-Minute Quadrangle, Georgia: Georgia Department of Natural Resources Open File Report 09-1, Plate 4
STOP 2: COOPER FURNACE AND OCONEE SUPERGROUP

Stop Leaders: Tim Chowns and Bill Witherspoon

This old furnace (Figure 1) is typical of the cold-blast, charcoal furnaces operated prior to and immediately after the Civil War. By about 1880 they were replaced by hot-blast furnaces utilizing first charcoal and later coke (Figure 2). It was constructed by the renowned ironmaster, Moses Stroup (1794-1877) who was also responsible for similar furnaces at Round Mountain, Cherokee Co., Tannehill, Tuscaloosa Co., and Oxmoor, near Birmingham, Alabama. The furnace was charged from the top via a trestle, with alternating layers of charcoal, limonitic ore and limestone, fired by blasts of air from a bellows powered by an overshot water wheel. Water was apparently carried by a wooden flume connected to the stream in Hurricane Creek rather than from the Etowah River. Molten iron was tapped at the base of the furnace and fed into ‘pigs’ on the sand floor of a casting shed while silica and other impurities combined with limestone to form a slag of calcium silicate. The three furnaces at Tannehill, in Alabama, have been restored and No 2 was fired up during the bicentennial of 1976 (Morris & White, 1997).

According to a trail guide prepared by the Corps of Engineers, Cooper Furnace supplied pig iron for the production of nails, spikes, rails, pots, tools, cannons and other related items and was the center for the once thriving town of Etowah founded in the late 1830’s by Jacob Stroup (1771-1846), father of Moses Stroup. In addition to the furnace there were spike and nail mills, a rolling mill, foundry and flour mill, as well as a hotel and homes, stretching for about a mile upstream from the iron works.

LUNCH STOP/BUSINESS MEETING

10.35/0.00  Return to bus, turn west on Old River Road
13.15/2.80  Turn right (north) on Highway 41 (Joe Frank Harris Pkwy)
17.15/4.00  Take Georgia 20E Exit toward Georgia 61 North
17.25/0.10  Merge onto Georgia 20 West
17.35/0.10 Turn right onto US411 north

20.55/3.20 Turn right onto Museum Drive, Clarion is on the left

END OF DAY 2.-END OF FIELDTRIP
## GEORGIA GEOLOGICAL SOCIETY GUIDEBOOKS

Paper copies of guidebooks can be ordered from: Georgia Geological Society
Department of Geosciences, University of West Georgia, Carrollton, GA 30118

Digital copies can be downloaded in Screen Optimized (SO) or Print Optimized (PO) PDF’s format from the GGS web site (http://www.westga.edu/~ggsweb)

<table>
<thead>
<tr>
<th>Date</th>
<th>Publication</th>
<th>Cost/Availability</th>
</tr>
</thead>
<tbody>
<tr>
<td>1967</td>
<td><strong>Geology of the Barnesville area and Towaliga Fault, Lamar County, Georgia.</strong> by W. H. Grant. Published by the Georgia Geological Survey.</td>
<td>Out SO</td>
</tr>
<tr>
<td>1968</td>
<td><strong>Late Tertiary stratigraphy of eastern Georgia.</strong> by S. M. Herrick and H. B. Counts. Printed at West Georgia College</td>
<td>Out</td>
</tr>
<tr>
<td>1969</td>
<td><strong>A guide to the stratigraphy of the Chickamauga Supergroup in its type area.</strong> by R. C. Milici and J. W. Smith Georgia Geologic Survey, Department of Natural Resources; published simultaneously as Report of Investigation no. 24, Tennessee Division of Geology.</td>
<td>Out</td>
</tr>
<tr>
<td>1970</td>
<td><strong>Stratigraphic and structural features between the Cartersville and Brevard Fault zones.</strong> by T. J. Crawford and J. H. Medlin. Georgia Geologic Survey, Department of Natural Resources.</td>
<td>Out</td>
</tr>
<tr>
<td>1971</td>
<td><strong>Norite intrusives in western Jasper County and eastern Monroe County, Georgia.</strong> by Robert H. Carpenter; Lithostratigraphy and biostratigraphy of the north central Georgia Coastal Plain. by Sam Pickering; The mining methods utilized by Freeport Kaolin Company at their mines near Gordon, Georgia. by J. H. Auvil. Published by the Georgia Geological Society - printed and distributed by the Georgia Geologic Survey.</td>
<td>Out</td>
</tr>
<tr>
<td>1973</td>
<td><strong>The Neogene of the Georgia Coast.</strong> by Robert W. Frey, ed. Publisher unknown.</td>
<td>Out</td>
</tr>
<tr>
<td>1974</td>
<td>(a) <strong>The Lake Chatuge Sill outlining the Brasstown antiform.</strong> by M. E. Hartley and H. M. Penley.  (b) <strong>An introduction to the Blue Ridge tectonic history of northeast Georgia.</strong> by R. D. Hatcher. Georgia Geologic Survey, Department of Natural Resources, Guidebook 13 and 13A.</td>
<td>Out</td>
</tr>
</tbody>
</table>

1 Print optimized guidebooks are available for more recent fieldtrips. These guidebooks are prepared from the original word processing files at 1200 dpi resolution.
<table>
<thead>
<tr>
<th>Date</th>
<th>Publication</th>
<th>Cost/Availability</th>
</tr>
</thead>
<tbody>
<tr>
<td>1976</td>
<td><em>Stratigraphy, structure, and seismicity in Slate Belt rocks along the Savannah River.</em> by T. M. Chowns. Georgia Geologic Survey, Department of Natural Resources, Guidebook 16.</td>
<td>Out</td>
</tr>
<tr>
<td>1977</td>
<td><em>Stratigraphy and economic geology of Cambrian and Ordovician rocks in Bartow and Polk Counties, Georgia.</em> by T. M. Chowns. Published by the Georgia Geological Society for the 12th Annual Meeting and Field Trip, printed at West Georgia College.</td>
<td>Out</td>
</tr>
<tr>
<td>1979</td>
<td><em>The Stratigraphy of the Barnwell Group of Georgia.</em> by Paul F. Huddlestun and John H. Hetrick. Reprinted 1988 by the Georgia Geological Society.</td>
<td>$10.00</td>
</tr>
<tr>
<td>1983</td>
<td><em>Geology of Paleozoic rocks in the vicinity of Rome, Georgia.</em> by T. M. Chowns, ed. Georgia Geological Society Guidebooks, vol. 3, no. 1.</td>
<td>$5.00</td>
</tr>
<tr>
<td>1985</td>
<td><em>Coastal processes and barrier island development, Jekyll Island, Georgia.</em> by V. J. Henry and W. J. Fritz, and Examination of the Altamaha Formation near Oak Park, Emanuel County, Georgia. by Paul F. Huddlestun. Georgia Geological Society Guidebooks, vol. 5, no. 1.</td>
<td>$10.00</td>
</tr>
<tr>
<td>Year</td>
<td>Title</td>
<td>Author(s)</td>
</tr>
<tr>
<td>------</td>
<td>----------------------------------------------------------------------</td>
<td>---------------------------------------------------------------------------</td>
</tr>
<tr>
<td>1986</td>
<td>Depositional systems of Pennsylvanian rocks in the Cumberland Plateau of southern Tennessee</td>
<td>H. G. Churnet and R. E. Burgenback. Field Trip no. 4, 1986 SEPM Annual Meeting. Georgia Geological Society Guidebooks, vol. 6, no. 4.</td>
</tr>
</tbody>
</table>
| 1986 | Gold and base metal mineralization host rocks in the Dahlonega and Carroll County Gold Belts, Georgia. | K. I. McConnell, J. M. German and C. E. Abrams, with a prologue by G. O. Allard. Georgia Geological Society Guidebooks, vol. 6, no. 5. | $10.00 | SO  
| 1988 | Geology of the Murphy Belt and related rocks, Georgia and North Carolina. | W. J. Fritz and T. E. La Tour, eds., Georgia Geological Society Guidebooks, vol. 8, no. 1. | $10.00 |  
| 1989 | The Geology of the East End of the Pine Mountain Window and Adjacent Piedmont, Central Georgia. | Robert J. Hooper and Robert D. Hatcher, Jr. (Fieldtrip No. 2, 1989 SE GSA, Atlanta) Georgia Geological Society Guidebooks, vol. 9, no. 2. | $5.00 | SO  

-59-
<table>
<thead>
<tr>
<th>Year</th>
<th>Title</th>
<th>Author/Editor</th>
<th>Price</th>
<th>Availability</th>
</tr>
</thead>
<tbody>
<tr>
<td>1997</td>
<td>Geology of the Piedmont in the vicinity of Athens and eastern metropolitan Atlanta area.</td>
<td>D.B. Wenner, ed., Georgia Geological Society Guidebooks, vol. 17, no. 1.</td>
<td>$10.00</td>
<td>SO</td>
</tr>
<tr>
<td>1999</td>
<td>An introduction to sequence stratigraphy: illustrations from the Valley and Ridge province in Georgia and Alabama.</td>
<td>T. M. Chowns, S. M. Holland, W. C. Elliott and The geology of Civil War battlefields in the Chattanooga and Atlanta Campaigns in the Valley and Ridge of Georgia.</td>
<td>$10.00</td>
<td>SO &amp; PO</td>
</tr>
<tr>
<td>2001</td>
<td>Across the Brevard Zone: The Chattahoochee Tunnel, Cobb County, Georgia.</td>
<td>R. L. Kath and T. J. Crawford, eds., Georgia Geological Society Guidebooks, vol. 21, no. 1.</td>
<td>$10.00</td>
<td>SO &amp; PO</td>
</tr>
<tr>
<td>2002</td>
<td>Geologic features of eastern Pickens, Dawson, and western Lumpkin Counties, Georgia.</td>
<td>J.O. Costello, ed., Georgia Geological Society Guidebooks, vol. 22, no. 1.</td>
<td>$10.00</td>
<td>SO</td>
</tr>
<tr>
<td>2003</td>
<td>Geology of the Americus Area,</td>
<td>M.D. Cocker and J.O. Costello, eds., Georgia Geological Society Guidebooks, vol. 23, no. 1.</td>
<td>$10.00</td>
<td>SO</td>
</tr>
<tr>
<td>2005</td>
<td>Investigations of Elberton Granite and Surrounding Rocks.</td>
<td>Mike Roden, Paul A. Schroeder, and Sam Swanson, eds., Georgia Geological Society Guidebooks, vol. 25, no. 1.</td>
<td>$10.00</td>
<td>PO</td>
</tr>
<tr>
<td>2006</td>
<td>Quaternary Stratigraphy and Depositional Environments; Jekyll Island and the Golden Isles Parkway.</td>
<td>Timothy M. Chowns. ed, Georgia Geological Society Guidebooks, vol. 26, no. 1</td>
<td>$10.00</td>
<td>PO</td>
</tr>
<tr>
<td>2008</td>
<td>The Emerson-Talladega Fault, the Great Smoky Fault, and adjacent Folding and Faulting: Geology and Historical Interpretations based on detailed Geologic Mapping in Polk and Bartow Counties, Georgia.</td>
<td>Randy L. Kath, ed., Georgia Geological Society Guidebooks, vol. 28, no. 1</td>
<td>$10.00</td>
<td>SO &amp; PO</td>
</tr>
<tr>
<td>2009</td>
<td>Fall Line Geology of East Georgia: with a special emphasis on the Upper Eocene.</td>
<td>Mack S. Duncan and Randy L. Kath, eds., Georgia Geological Society Guidebooks, vol. 29, no. 1</td>
<td>$10.00</td>
<td>SO &amp; PO</td>
</tr>
</tbody>
</table>
2010  **Environmental Geology and Hydrogeology**: by Ronald Wallace, ed., Georgia Geological Society Guidebooks, vol. 30, no. 1  $10.00


2012  **Aggregate Resources and the Permo-Jurassic Geology of the southeastern Georgia Piedmont**: by R. Scott Harris and Jerry M. German, eds., Georgia Geological Society Guidebooks, vol. 32, no. 1.  $10.00  SO

2013  **The Dahlonega Wine and Gold District: Geology and Terroir of Viticulture in Northeastern Georgia**: by Paul Schroeder, Joseph Forrest, and Jerry German, eds., Georgia Geological Society Guidebooks, vol. 33, no. 1.  $10.00  SO

2015  **Origin of ore deposits in the Cartersville Mining District & Stratigraphic and Kinematic evidence for the separation of the Cartersville-Great Smoky and Emerson-Talladega Faults**: by Randy Kath and Timothy Chowns, eds., Georgia Geological Society Guidebooks, vol. 34, no. 1.  $10.00  PO

**Out** = paper copy is out-of-print  
**SO** = Screen Optimized (scanned 400 dpi)  
**PO** = Print Optimized (1200 dpi)

TOTAL ______________________

*Note: The Georgia Geological Society Guidebooks series starts with the 1981 issue. Even though the 1981 to 1987 issues contain no numbers on the first printing; they are considered part of the series. Guidebooks for the 1966-1980 meetings were published and distributed mostly by the Georgia Geologic Survey and variously referenced (Fritz, Power and Cramer, GGS Guidebooks, vol. 8, no. 1, p. 1-2). Even though these older guidebooks are part of the set of publications associated with meetings of the Georgia Geological Society, they are not considered part of the Georgia Geological Society Guidebook series.

To order PAPER copies of guidebooks or to get more information contact: 
Dr. Timothy M. Chowns  
Georgia Geological Society  
Dept. of Geosciences  
University of West Georgia  
Carrollton, GA 30118  
Phone: 678-839-4052  Fax: 678-839-4071  email: tchowns@westga.edu