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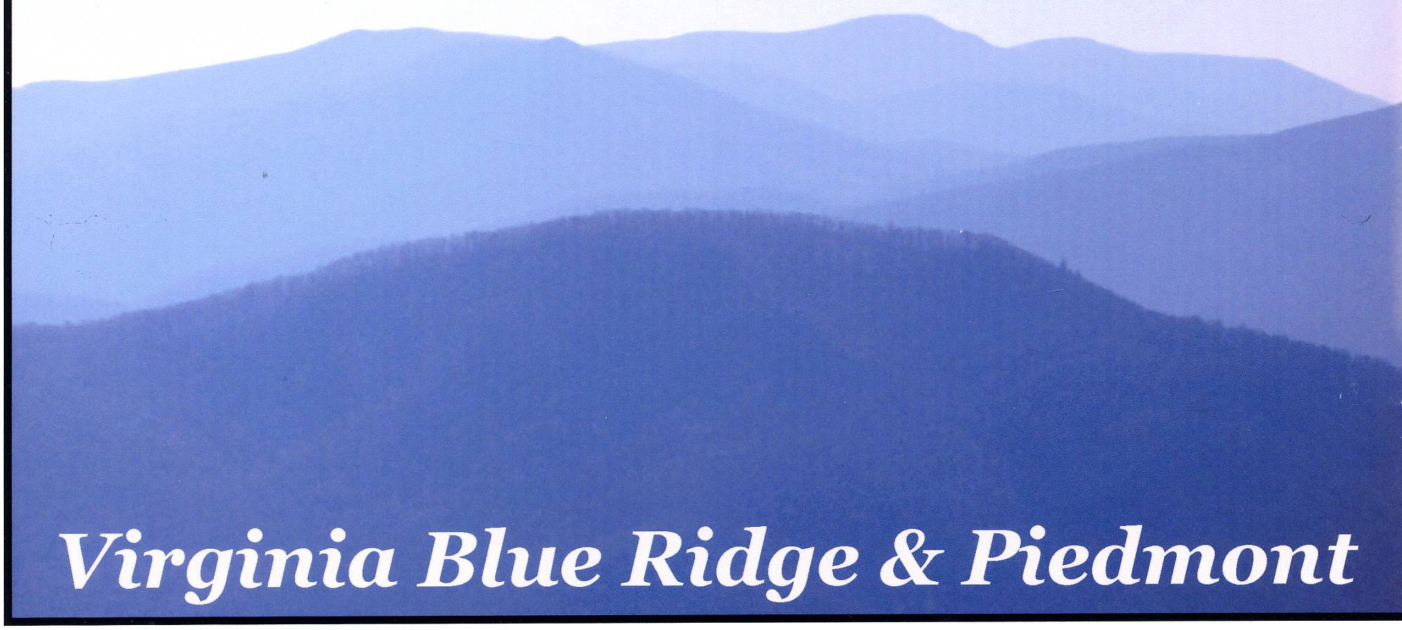
Kinematics & Vorticity of High-Strain Zones



Virginia Blue Ridge & Piedmont

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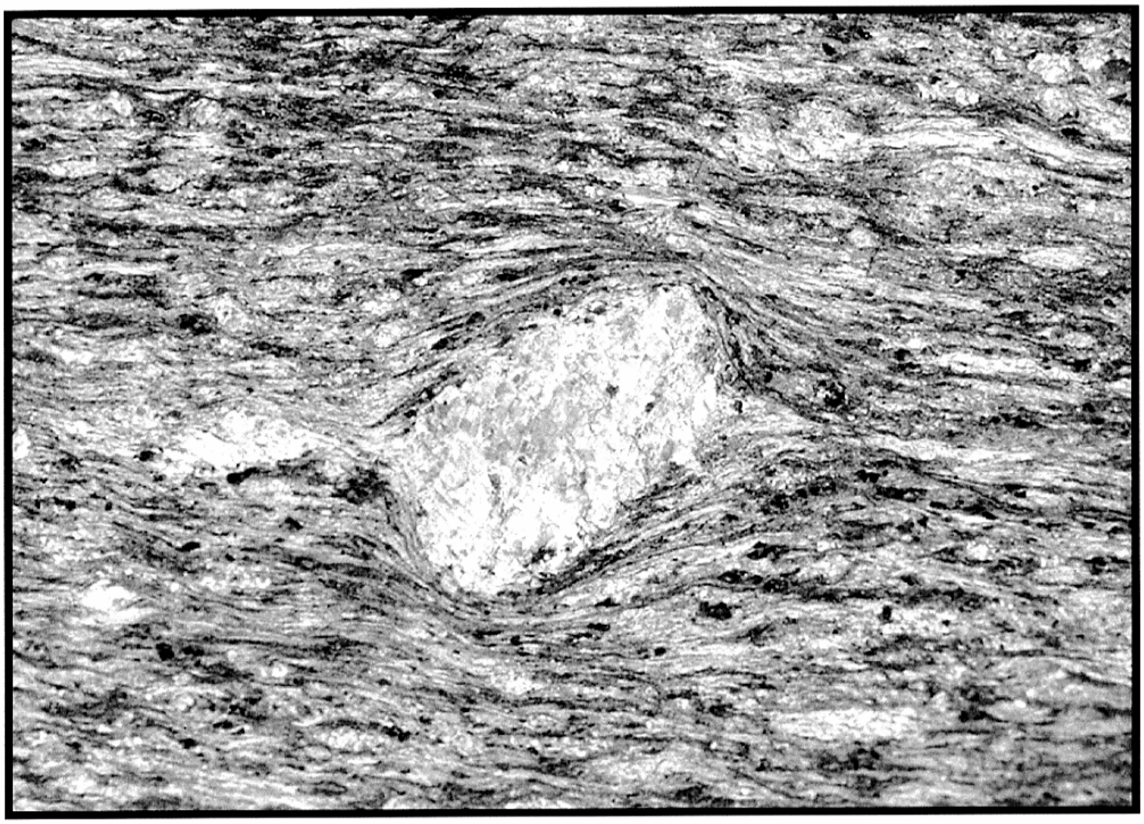
Geological Society of America Field Forum
April 16-21st 2002



Back-rotated feldspar porphyroclast in charnockitic mylonite from the Quaker Run high-strain zone, Blue Ridge province. Field of view ~9 x 15 cm.

"Ignorance is preferable to error; and he is less remote from the truth who believes nothing, than he who believes what is wrong."

Thomas Jefferson, *Notes on the State of Virginia* (1787)

Purpose

High-strain zones are common features in orogenic belts and have long generated spirited debates as to their movement history and tectonic significance. A number of notable advances have recently been made towards quantitatively understanding the nature of flow during deformation, but very few studies have applied these methods to naturally deformed rocks. The purpose of this field forum is to examine a number of high-strain zones and deformed rocks in the Blue Ridge and Piedmont provinces of the Virginia Appalachians and discuss what information can be gained into the kinematic history of these rocks. We hope this field forum provides an opportunity to discuss contemporary research in the context of what *can* and *cannot* be learned about the deformation history of naturally deformed rocks.

Questions

1) Can structures in natural high-strain zones be used to meaningfully characterize the finite vorticity and progressive vorticity changes during deformation?

2) How does the recognition of triclinic symmetries in high-strain zones influence kinematic and tectonic interpretations?

3) How many measurements/estimates of strain and vorticity are needed to characterize the bulk flow of a high-strain zone? How do these quantities vary within a single high-strain zone?

4) Is the deformation path (including strain symmetry, vorticity, three-dimensional strain, and volume change) influenced by the specific tectonic environment in which material is deformed?

5) What is the proper context in which to describe deformation as transpressional?

6) On what should future research focus?

Itinerary

Tuesday, April 16th: Travel to Graves Mountain Lodge (540-923-4231) in Madison County, Virginia (30 miles north of Charlottesville). Arrive late afternoon with dinner at 6:30 p.m. Orientation and introductory presentations in the evening.

Wednesday, April 17th: Breakfast at 8 a.m., depart for eastern Piedmont by 8:30 a.m. (~70 miles). Examine mylonitic rocks in the Hylas high-strain zone, mylonitic gneiss and deformed pegmatite in the Goochland terrane, L-tectonites and deformed pebble metaconglomerates in the Chopawamsic terrane, and the Mountain Run fault zone. Return to Graves Mountain Lodge by 6:30 p.m. for dinner. After dinner informal discussions on high-strain zones: past models to current thinking.

Thursday, April 18th: Breakfast at 8 a.m., the entire day will be spent examining Mesoproterozoic basement rocks and high-strain zones exposed in the Blue Ridge anticlinorium in Madison County. In the morning we will hike 2 km to a debris flow scar that exposes the Upper Kinsey Run high-strain zone. The afternoon will be spent examining other mylonitic rocks exposed in debris flow scars and roadside quarries. Return to Graves Mountain Lodge by 6:30 p.m. for dinner. After dinner informal discussions: critically evaluating field studies and the outlook for future research.

Friday, April 19th: Breakfast at 8 a.m., the entire day will be spent examining Mesoproterozoic basement rocks, Neoproterozoic metasedimentary rocks, and high-strain zones exposed in the Blue Ridge anticlinorium in Madison County. The afternoon will be spent on a large exposure in the Garth Run high-strain zone, a ~100 meter thick zone of heterogeneously deformed mylonitic rock. Depart from Graves Mountain Lodge, dinner in Charlottesville, and lodging at the Inn at Afton (540-942-5201).

Saturday, April 20th: Breakfast at 7 a.m., the morning will be spent examining exposures in the western Blue Ridge and the Rockfish Valley high-strain zone. In the afternoon we will examine deformed metaconglomerates exposed at the Blue Ridge basement-cover contact, basement lithologies, and the Lawhorne Mill high-strain zone in Nelson County (~30 miles SW of Charlottesville). Return to the Inn at Afton for dinner and summary presentations.

Sunday, April 21st: Travel to the Charlottesville-Albemarle airport for flights.

Geologic Setting

Blue Ridge province

The Blue Ridge province forms the major basement massif in the central and southern Appalachians; separating the fold and thrust belt of the Valley and Ridge from the orogenic hinterland exposed in the Piedmont to the southeast (Fig. 1). In central and northern Virginia, the Blue Ridge forms an anticlinorium with Mesoproterozoic basement rock in the core and Neoproterozoic to early Paleozoic cover rock on the flanks and in fault-bounded inliers (Figs. 2 and 3). The Blue Ridge basement is considered ancestral North American crust (Laurentian) upon which a Neoproterozoic rift sequence and early Paleozoic passive margin sequence were deposited (Rankin et al., 1989). The Blue Ridge is allochthonous and was thrust to the northwest over early Paleozoic rocks of the Valley and Ridge during the late Paleozoic Alleghanian orogeny (Evans, 1989) (Fig. 2). The structural geometry of the Blue Ridge province is that of an imbricated stack of basement thrust sheets that experienced 40-60% shortening during Paleozoic deformation (Evans, 1989; Bailey, 1994).

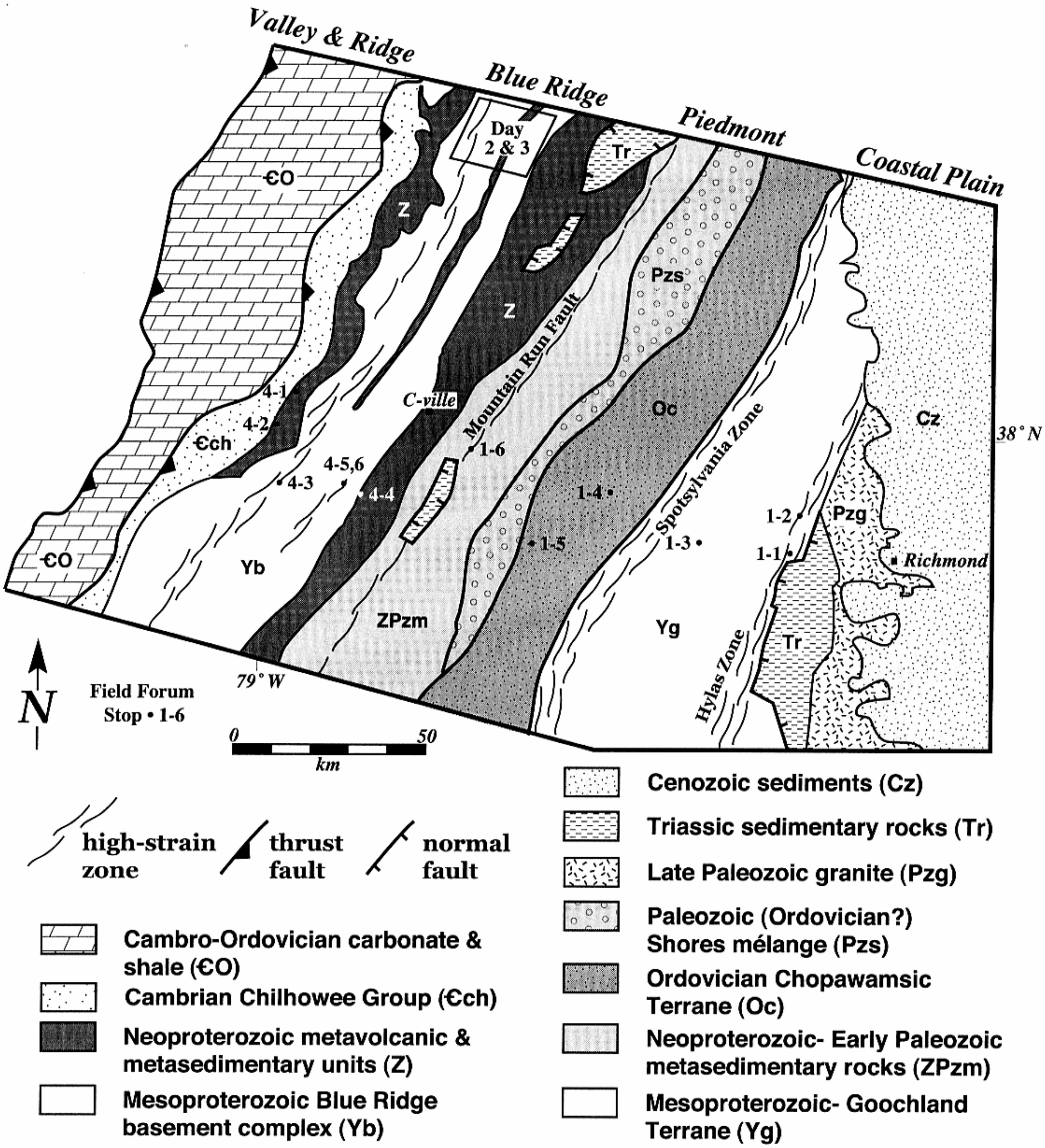


Figure 1. Generalized geologic map/terrane map of the central Virginia Appalachians. Modified from Virginia Division of Mineral Resources (1993) and other sources.

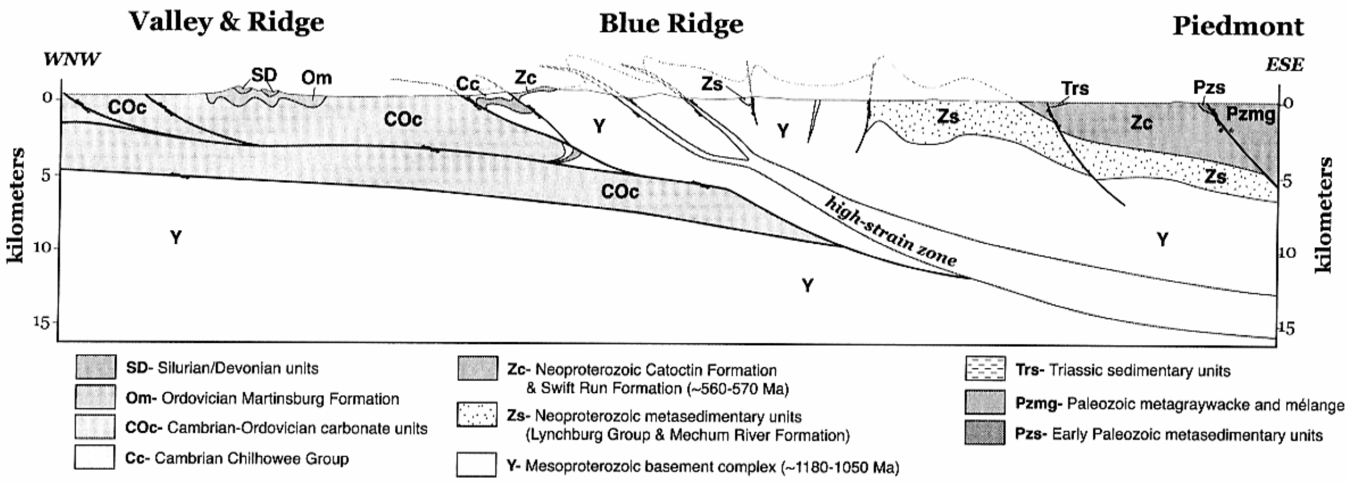


Figure 2. Generalized geologic cross section of Valley & Ridge, Blue Ridge, and westernmost Piedmont. Modified from Evans, 1989; Lampshire et. al. 1993; and Bailey, 1994.

The oldest rocks exposed in the core of the anticlinorium are layered gneisses (mostly orthogneisses) that were intruded by a series of 1180 to 1050 Ma charnockitic, leucogranitic, and anorthositic plutons during the Grenvillian event (Fig. 3) (Sinha and Bartholomew, 1984; Herz and Force, 1987; Aleinikoff and others, 2000; Tollo and Aleinikoff, 2001). Grenvillian deformation occurred under upper amphibolite to granulite facies conditions and recent geochronology brackets the age of a major deformation event between 1060-1050 Ma (Tollo and Aleinikoff, 2001).

A distinct group of Neoproterozoic A-type granitoid plutons intrude the Blue Ridge basement (Fig. 3). These plutons include the 735-705 Ma Robertson River Igneous Suite in northern Virginia (Tollo and Aleinikoff, 1996) and the Rockfish River, 706 ± 4 Ma Polly Wright Cove, and Mobley Mountain plutons in central Virginia (Sinha and Bartholomew, 1984; Bailey and Tollo, 1998; Herz and Force, 1987). These granites were emplaced as a series of dike-like plutons during active crustal extension associated with the early stages of Iapetan rifting (Bailey and Tollo, 1998).

Both Mesoproterozoic basement units and Neoproterozoic plutons are unconformably overlain by clastic metasedimentary rocks of the Lynchburg Group, Mechum River Formation, and Swift Run Formation (Fig. 3). These units are arkosic in character, range from non-marine alluvial conglomerates to deep-water distal turbidites, and are interpreted to be rift-related (Wehr, 1983). Some of these deposits record glaciogenic sedimentation associated with Neoproterozoic ice house conditions (Wehr, 1986; Bailey and Peters, 1998). On the east limb of the Blue Ridge anticlinorium the Lynchburg Group is ~5 kilometers thick (Fig. 2). The Mechum River Formation crops out in a narrow (<2 km wide) belt in the center of the Blue Ridge anticlinorium. In the western Blue Ridge Neoproterozoic units are locally absent and were deposited in basins associated with rift-related normal faults (Bailey et al., 2002). The areal distribution of Neoproterozoic sedimentary units across the Blue Ridge province suggests that it lay on the western edge (present day geography) of a rifted continental margin, and that the Iapetus Ocean opened to the east (Wehr and Glover, 1985).

Neoproterozoic mafic to ultramafic dikes and sills of uncertain affinity intrude the basement complex and Neoproterozoic metasedimentary units in the eastern Blue Ridge (Fig. 3).

On both limbs of the Blue Ridge anticlinorium, the ~570 Ma Catoctin metabasalts (Badger and Sinha, 1988; Aleinikoff and others, 1995) overlie the basement complex and Neoproterozoic metasedimentary units (Fig. 2 and 3). Catoctin basalts were extruded over a large region (>4000 sq. km), subareally in the western Blue Ridge and subaqeously in the eastern Blue Ridge. On the western flank of the Blue Ridge, Catoctin metabasalts are overlain by the Neoproterozoic to Early Cambrian Chilhowee Group a sequence of siliciclastic rocks that records the Iapetan rift to drift transition (Fig. 3) (Simpson and Erikkson, 1989). In the eastern Blue Ridge the Catoctin Formation is overlain by a sequence of low grade metasedimentary rocks interpreted to have been deposited under marine conditions in the latest Neoproterozoic to Early Paleozoic (Fig. 3) (Evans and Milici, 1994). By the early Cambrian, the Iapetus Ocean had opened to the east and a westward transgression occurred on the Laurentian margin. Throughout the early Paleozoic a thick passive margin sequence was deposited along the Laurentian margin and these rocks are preserved today as Cambro-Ordovician carbonates in the Valley & Ridge province to the west (Figs. 1-3).

Rocks in the Blue Ridge were metamorphosed to the greenschist facies and brought to the surface during an Ordovician orogenic event (Taconic orogeny) as illustrated by clasts of metamorphosed basement and Chilhowee Group quartzite clasts in late Ordovician conglomerates exposed in the Valley and Ridge (Karpa, 1974). The absolute age of Paleozoic deformation in the Virginia Blue Ridge is not precisely known, however the available metamorphic cooling ages suggest greenschist facies fabrics in both the basement and cover sequence developed during this Ordovician event (~450-425 Ma) (Bartholomew et al., 1991; Evans, 1991). However more recent Ar-Ar work indicates late Paleozoic (~320-290 Ma) cooling ages for the northern Virginia Blue Ridge (Kunk and Burton, 1999).

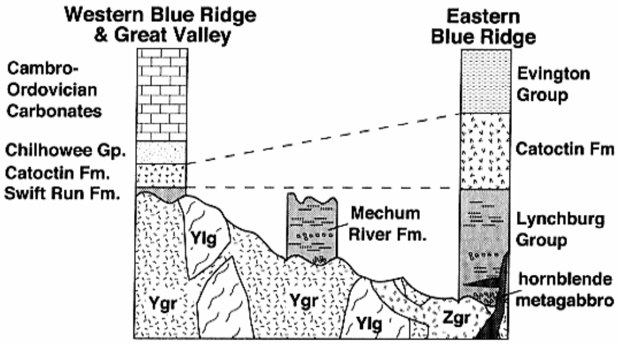


Figure 3. Generalized stratigraphy of the central Virginia Blue Ridge province. Ylg- Mesoproterozoic layered gneiss, Ygr- Mesoproterozoic granitoid intrusions, Zgr- Neoproterozoic granitic plutons.

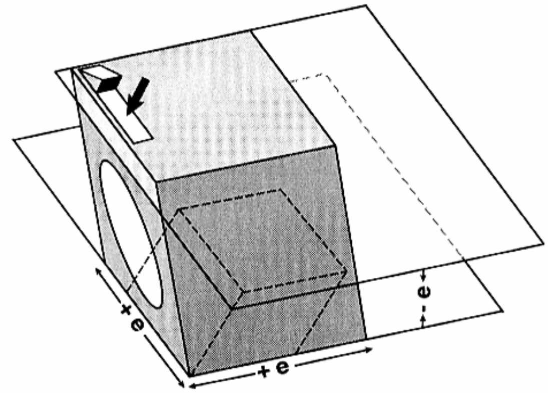


Figure 4. Idealized finite deformation for Blue Ridge high-strain zones characterized by general shear, flattening strain, and a weak triclinic symmetry. The overall tectonic transport records reverse displacement to the northwest.

Basement rocks in the Blue Ridge are cut by a series of anastomosing high-strain zones. The significance of these zones was first recognized by Mitra (1977, 1979) who referred to them as ductile deformation zones (ddz's). Major high-strain zones (Rockfish Valley, White Hall, Quaker Run, and Sperryville) are 0.5 to 3 km wide and extend 50-150 km parallel to the regional trend of the Blue Ridge. In central Virginia Bartholomew et al. (1981) first recognized the Rockfish Valley zone to be a major tectonic boundary separating basement massifs of distinctly different character. Different workers have interpreted the Rockfish Valley zone to have experienced normal, reverse, and strike-slip movement.

These zones are characterized by greenschist facies mylonitic rocks that dip to the southeast. Mineral elongation lineations commonly plunge downdip to obliquely downdip. Kinematic analysis reveals that these rocks record top-to-the northwest reverse sense-of-shear. Deformation is heterogeneous and penetrative foliations diminish away from the high-strain zones. Most of these zones experienced general shear (W_m 0.2-0.8), flattening strain, and are characterized by a weak triclinic symmetry (Fig. 4). Displacement, based on shear strain estimates, across these high-strain zones ranges from 0.5 to <3 km. The absolute age of movement on Blue Ridge high-strain zones is not precisely known. Field relations indicate that mylonitic high-strain zones are cut by brittle thrusts of Alleghanian age.

Pre-Paleozoic deformation fabrics and mylonitic high-strain zones occur in the southeastern Blue Ridge (Bailey and Simpson, 1993; Knight and Bailey, 1999). These amphibolite to greenschist facies high-strain zones record down-to-the-southeast extensional movement. Many of these extensional high-strain zones occur within and close to Neoproterozoic plutons. In Nelson County, ~40 km southwest of Charlottesville, the Lawhorne Mill high-strain zone (Stop 4-6) forms a 15 km-long zone of steeply dipping mylonitic rock interpreted to be a complex, reactivated Neoproterozoic high-strain zone (Bailey and Simpson, 1993). In Madison County, ~40 km northeast of Charlottesville, the White Oak Run high-strain zone records extensional deformation under amphibolite facies conditions (Knight and Bailey, 1999). Ar-Ar ages for recrystallized hornblende in these mylonites yield disturbed spectra with integrated ages of ~650 to ~ 400 Ma.

Major unresolved geologic problems in the Blue Ridge province include 1) the age of ductile thrusting, 2) the age and significance of mafic to ultramafic intrusive rocks, and 3) the nature of the basement-cover contact in the eastern Blue Ridge (unconformity or tectonic contact?).

Piedmont province

The Piedmont province is characterized by a gently rolling landscape underlain by igneous and metamorphic rocks that form the internal core of the Appalachian orogen. Rocks are strongly weathered in the Piedmont's humid climate and bedrock is generally buried under a thick (2-20 m) blanket of saprolite. A number of distinct terranes have been recognized in the Piedmont including exotic terranes accreted to Laurentia during the Paleozoic. From east to west these terranes include the Goochland, Chopawamsic, Western Piedmont, and Blue Ridge (Laurentia) (Figs. 1 and 5). Early Mesozoic crustal extension associated with the opening of the Atlantic Ocean produced a series of rift-basins filled with non-marine sedimentary rocks and a suite of N-NW striking diabase dikes.

The 330 ± 8 Ma Petersburg granite forms a large pluton in the easternmost Piedmont (Fig. 1) (Wright et al., 1975). The Petersburg Granite was juxtaposed against Mesoproterozoic rocks of the Goochland terrane along the Hylas Zone, a ductile high-strain zone with right-lateral offset (Bobyarchick, 1981; Bobyarchick and Glover, 1978; Gates and Glover, 1989) (Fig. 5). The Hylas Zone records amphibolite to greenschist facies deformation and Ar-Ar cooling ages of 300 to 240 Ma (Gates and Glover, 1989). The Hylas Zone was reactivated as a brittle structure during the formation of the Triassic Richmond basin.

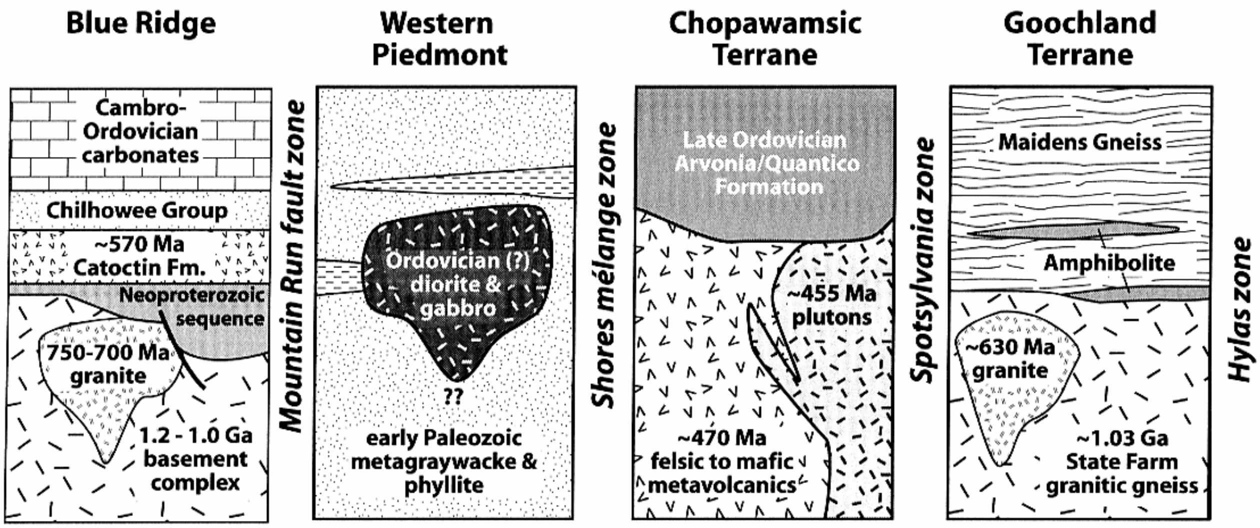


Figure 5. Generalized stratigraphy of terranes in the central Virginia Piedmont and Blue Ridge provinces.

The Goochland terrane is composed of multi-deformed and metamorphosed gneiss, amphibolite, granite, and anorthosite (Fig. 5). U-Pb zircon ages for the State Farm granitic gneiss and Montpelier anorthosite yield Mesoproterozoic ages of 1050-1020 Ma (Aleinikoff et al., 1996; Owens and Tucker, 1999). A suite of small A-type granitoid plutons with U-Pb zircon ages of ~630 Ma intrude the State Farm granitic gneiss (Owens and Tucker, 2000). The State Farm gneiss crops out in a series of domes that are overlain by amphibolite and the heterogeneous Maidens gneiss (Poland, 1976). The Maidens is dominantly a pelitic unit with some granitic gneiss. Rocks of the Goochland terrane experienced an early granulite facies metamorphic event that was overprinted by a later amphibolite facies event (Farrar, 1984; Farrar and Owens, 2001). Farrar (1984) interprets the early granulite facies metamorphism as Mesoproterozoic and the amphibolite facies event as Alleghanian (~300- 250 Ma). The origin of the Goochland terrane is unclear: the Mesoproterozoic rocks and A-type Neoproterozoic granitoids are similar to Laurentian basement in the Blue Ridge (Farrar, 1984; Glover, 1989; Aleinikoff et. al., 1996), however, others workers suggest the Goochland terrane may be of peri-Gondwanan affinity and was accreted to Laurentia during the Appalachian orogen (Rankin et al, 1989; Hibbard and Samson, 1995).

In central Virginia the Goochland terrane is bounded on the west by the Spotsylvania zone. This feature was first recognized as a geophysical lineament separating rocks with distinctly different magnetic and radiometric signatures. Francis (2001) recognized a wide zone of mylonitic rocks along the James River, that strike NE-SW, dip moderately to the SE, and carry a gently NE plunging elongation lineation. These mylonitic rocks experienced general shear deformation consistent with dextral transpression under upper greenschist facies to amphibolite facies conditions. To the southwest the Spotsylvania Zone may link with the Hyco Zone, part of the Central Piedmont Shear Zone, a major tectonic boundary in the southern Appalachians (Hibbard et. al., 1998).

The Chopawamsic terrane is characterized by a suite of Ordovician (470- 450 Ma) felsic to mafic metavolcanic rocks and their intrusive equivalents (Fig. 5) (Coler et al., 2000). Metavolcanic rocks are both interlayered with and unconformably overlain by a sequence of metasedimentary rocks (Arvonia and Quantico Formations). At low metamorphic grades these metasedimentary rocks preserve late Ordovician fossils. The Chopawamsic terrane is interpreted to be an Ordovician arc complex developed on continental crust outboard of Laurentia that was accreted during the Ordovician Taconic orogeny (Glover, 1989).

The interpreted suture between the Chopawamsic arc and Laurentia is a belt of strongly deformed rock (locally known as the Shores mélange) that contains isolated blocks or pods of mafic and ultramafic igneous rock in a highly contorted matrix of deformed and metamorphosed graywacke (Brown, 1985). West of the Shores mélange the Piedmont is underlain by a belt of low-grade metagraywacke and phyllite, known as the western Piedmont or Hardware terrane (Fig. 5) (Evans and Milici, 1994). Mafic to ultramafic bodies are common in this belt. This package is considered early Paleozoic and thought to have been deposited in a marine environment on the edge of Laurentia (Glover, 1989).

The Mountain Run fault zone forms the boundary between the Piedmont and Blue Ridge provinces. This narrow fault zone forms a linear topographic low in central Virginia and rocks are not well-exposed. Evans and Milici (1994) interpret this zone to be a Paleozoic thrust fault, however Bobyarychick (1999) notes evidence for dextral movement across this zone. Rocks on both sides of this zone are at the greenschist facies suggesting that thrust displacement is limited. To the northeast, the Mountain Run fault zone is cut off by the Culpeper Mesozoic basin, and to the southwest it links up with the Bowens Creek fault, a major structure in the southwestern Virginia Piedmont.

Day 1

Drive from Graves Mountain Lodge to the eastern Piedmont (~90 minutes)

Stop 1-1

Petersburg Granite in the Hylas Zone & Triassic Faults at the Luck Stone Boscobel Quarry

Stop 1-2

Mylonitic rocks in the Hylas Zone at the Vulcan Royal Stone Quarry

Stop 1-3

Maidens Gneiss, Hidden Rock Park, Goochland

Strongly foliated biotite-garnet bearing gneiss (plag, qtz, bt, grt) is intruded by a number of pegmatitic dikes at this exposure (Fig. 6). This rock has been mapped as the Maidens Gneiss in the Goochland terrane (Farrar, 1984; Virginia Division of Mineral Resources, 1993). The foliation has been folded into a series of recumbent antiforms and synforms. Some pegmatitic dikes are discordant to the foliation whereas other pegmatitic dikes are folded and boudinaged. These dikes serve as minimum strain markers and record sectional strains of ~15 to 20:1. Elongation lineations, where discernible, plunge gently to both the northeast and southeast. Kinematic indicators may be compatible with dextral shear, but do these rocks record dextral transpression and/or multiple deformation events? The western limit of the Spotsylvania zone is 15 km to the west, but our reconnaissance work indicates a broad zone of highly strained rocks in the western Goochland terrane.

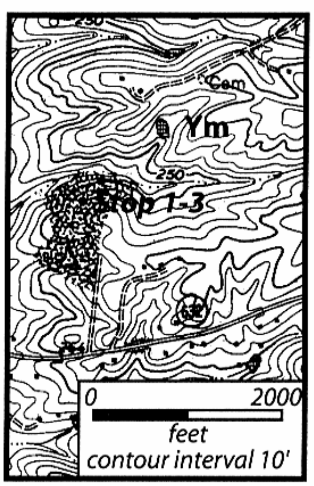


Figure 6. Stop 1-3 at Hidden Rock Park (former landfill) approximately 2 miles east of Goochland Courthouse. Perkinsville 7.5' quadrangle. Ym-Madiens gneiss.

Stop 1-4

L-tectonites in the Columbia granodioritic gneiss, Cowherd Quarry, Columbia

Granodioritic gneiss of the Columbia pluton is exposed in this old roadside quarry (Fig. 7). The 457± 7 Ma (Sinha, pers. comm., 2002) Columbia pluton ranges from a granite to a quartz diorite and intrudes metavolcanic rocks of the Chopawamsic Formation (Smith et al., 1964; Goodman et al., 2001). The Columbia pluton is part of a suite of Ordovician plutons that record magmatic activity associated with the Taconic orogeny in the Piedmont.

At the Cowherd quarry the granodioritic gneiss contains plagioclase, quartz, biotite, K-feldspar, and epidote with minor amounts of garnet, muscovite, and opaque minerals. Feldspar and quartz microstructures are consistent with recrystallization under amphibolite facies conditions. This rock forms a distinctive L-tectonite. The penetrative fabric is defined by aligned biotite and quartz aggregates. The lineation plunges ~25° towards ~050° and a very weak foliation strikes ~055° and dips ~80° northwest (Fig. 8). Poles to biotite cleavage form a strong great circle girdle normal to the lineation (Fig. 8). Three dimensional quartz fabrics in the *XZ* section range from 2.7-3.4 with K-values of 4-12 (strongly constrictional). Locally the L-tectonites are restricted to the nose of a map-scale northeast plunging synform. Three dimensional quartz fabrics in the Columbia pluton exhibit a complete range from L, L/S, and S tectonites.

Did this L-tectonite develop during a single progressive deformation or is it the result of two superposed deformations with orthogonal shortening directions? There is no microstructural evidence for two deformational/metamorphic events in these rocks. At the pluton scale the foliation is folded into a series of asymmetric northeast plunging folds that parallel the elongation lineation, requiring two deformation events. Although the quartz fabrics do not, *in a strict sense*, record the finite strain they are a measure /approximation of the overall strain. Here at the Cowherd quarry and at many locations in the central Piedmont material was apparently elongated in an orogen-parallel direction.

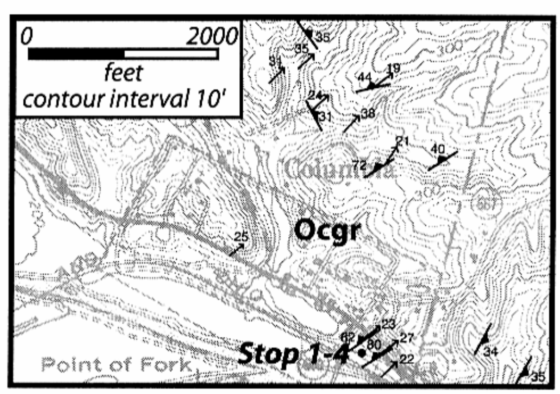


Figure 7. Stop 1-4 at Cowherd Quarry, Columbia. Columbia 7.5' quadrangle. Ocgr- Granodioritic gneiss of the Columbia pluton.

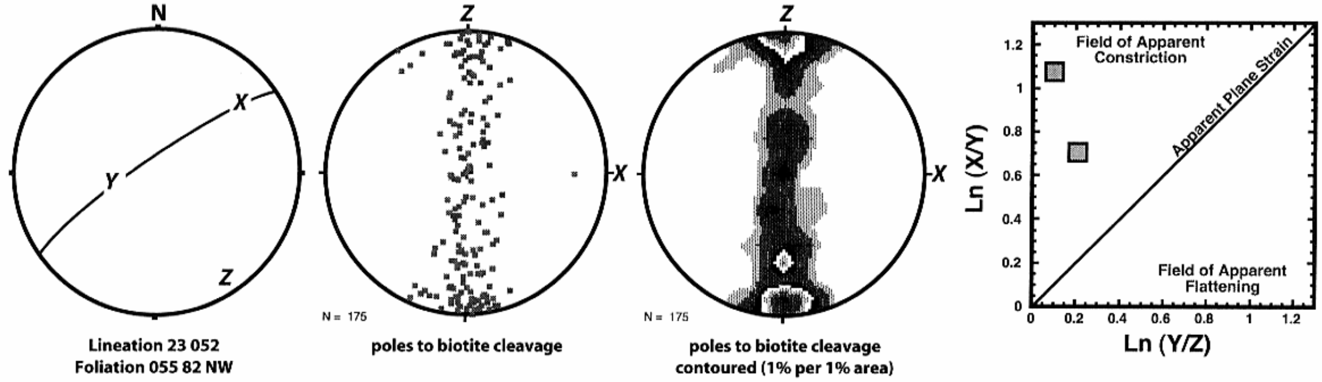


Figure 8. Synoptic diagram for fabric elements in the Columbia granodioritic gneiss at the Cowherd quarry, poles and contoured poles to biotite cleavage, and logarithmic Flinn diagram for quartz grain shapes (n = 40-64 per sample).

The Columbia pluton is unconformably overlain by the late Ordovician Arvonia Formation and thus was tectonically exhumed relatively rapidly during or after the Taconic orogeny. The existing K-Ar ages for micas in the gneiss yield cooling ages of ~300 Ma; consistent with Alleghanian metamorphism/deformation (Smith et al., 1964).

Stop 1-5 (time permitting)
Buffards metaconglomerate, near Penlan

The Buffards metaconglomerate crops out in the medial zone of the Arvonia synform (Brown, 1969). The metaconglomerate is a distinctive rock with elongate clasts of white quartz and a maroon-gray phyllite in a fine-grained matrix of quartz, chlorite, and muscovite (Fig. 9). The rock carries a northeast-striking penetrative foliation that dips steeply. At this location the rock is a LS-tectonite with a moderately developed gently plunging lineation defined by elongate pebbles. Phyllite clasts with aspect ratios of 20:1 are common, while the aspect ratio in quartz clasts is <5:1.

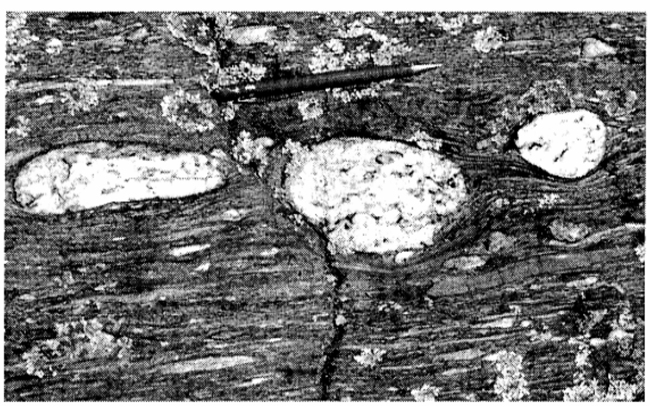


Figure 9. Buffards metaconglomerate with elongate quartz and phyllite pebbles.

Stop 1-6 (time permitting)
Everona metalimestone near the Mountain Run fault zone

The Everona metalimestone is exposed in this old roadside quarry in the Mountain Run fault zone. The Everona is a fine-grained laminated limestone and at this exposure layering dips to the southeast. In spite of the apparent structural simplicity at this outcrop, Evans and Milici (1994) noted intrafolial isoclinal folds on blocks and argued that these rocks are highly tectonized and exposed on the northwest overturned limb of an anticline.

The Everona metalimestone contains Cambro-Ordovician fossils and has been correlated with carbonates in the Valley & Ridge. Evans (1994) placed the Mountain Run fault zone approximately 0.5 km to the southeast of this quarry and interprets the fault to be a thrust that has been folded by later contraction.

Return to Graves Mountain Lodge (~ 80 minutes)

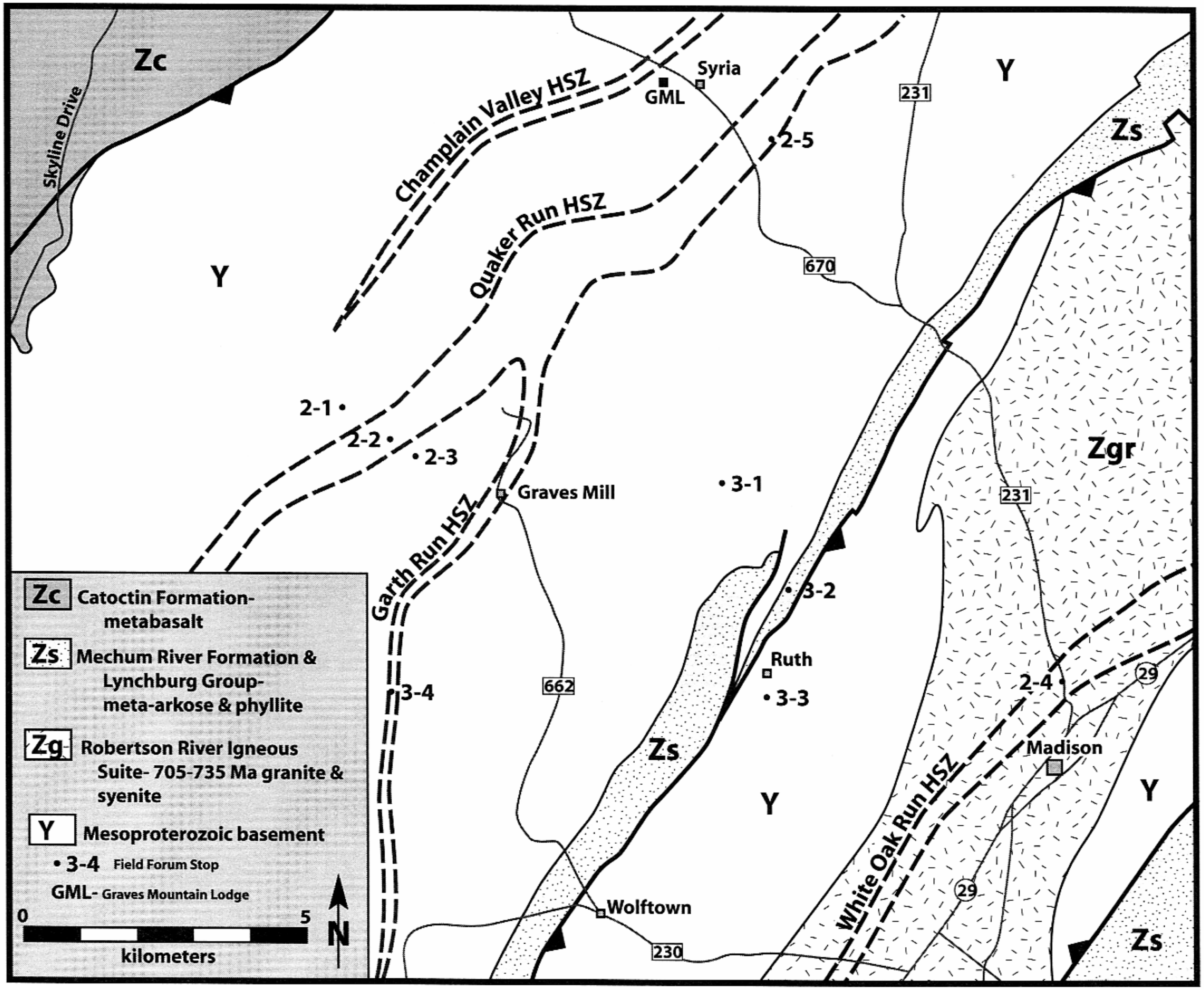


Figure 10. Generalized geologic map of the Madison area, Blue Ridge province.

Day 2

Stop 2-1

Kinsey Run high-strain zone

We will hike about 2 km (mostly uphill!) along a logging trail to reach the upper part of Kinsey Run (Fig. 11). At Kinsey Run three diabase dikes are dramatically exposed in a debris slide scar. This dikes intrude medium- to coarse-grained charnockite, strike ~350°, and are subvertical. The contacts are well exposed and have chilled margins. The diabase is composed of augite and plagioclase with minor amounts of magnetite and quartz. These dikes are part of a widespread suite of Jurassic (~200 Ma) igneous bodies that intruded Blue Ridge and Piedmont country rocks during early Mesozoic rifting.

A ~30 m wide high-strain zone is exposed in the debris flow channel above the diabase dikes. The Kinsey Run high-strain zone strikes ~080° and dips steeply to the northwest (Fig. 12). The zone is bounded by charnockite that passes into protomylonite and mylonite further into the high-strain zone. Two deformed metabasalt dikes occur within the zone and strike subparallel to the zone boundary (Fig. 12). Epidotized leucocharnockite dikes also occur in the mylonitic basement. Foliation dips steeply to both the northwest and southeast and a down-dip mineral elongation lineation occurs at a few locations (Fig. 12 and 13).

Asymmetric kinematic indicators generally display a reverse sense of shear (top-to-the southeast, Fig. 12) on foliation normal/lineation parallel sections (*XZ*). On sections normal to both the foliation and lineation (*YZ*) weakly asymmetric structures with an overall sinistral (strike-parallel) sense of shear occur (when viewed looking down the elongation lineation). The geometry of the fabric elements and kinematic indicators from the upper Kinsey Run high-strain zone are consistent with a weak triclinic deformation symmetry.

Undeformed charnockite is composed of 50% alkali feldspar (commonly perthitic), 20% quartz, 10-15% orthopyroxene, and 5-10% plagioclase with minor amounts of biotite, epidote, apatite, and titanite. Quartz grains (2-7 mm) are bleb-like and display no grain shaped preferred orientation. Orthopyroxene grains exhibit some alteration to fine-grained uralitic hornblende and secondary biotite that is localized along grain boundaries. Protomylonite and mylonite are composed of 30-50% quartz, 10-25% fine-grained muscovite, 10-20% alkali feldspar, 10-15% epidote, and 10% biotite. Feldspars are extensively fractured with quartz filled cracks and muscovite mantles along grain boundaries. Quartz grains form monocrystalline lenses and ribbons with a moderate to strong crystallographic preferred orientation.

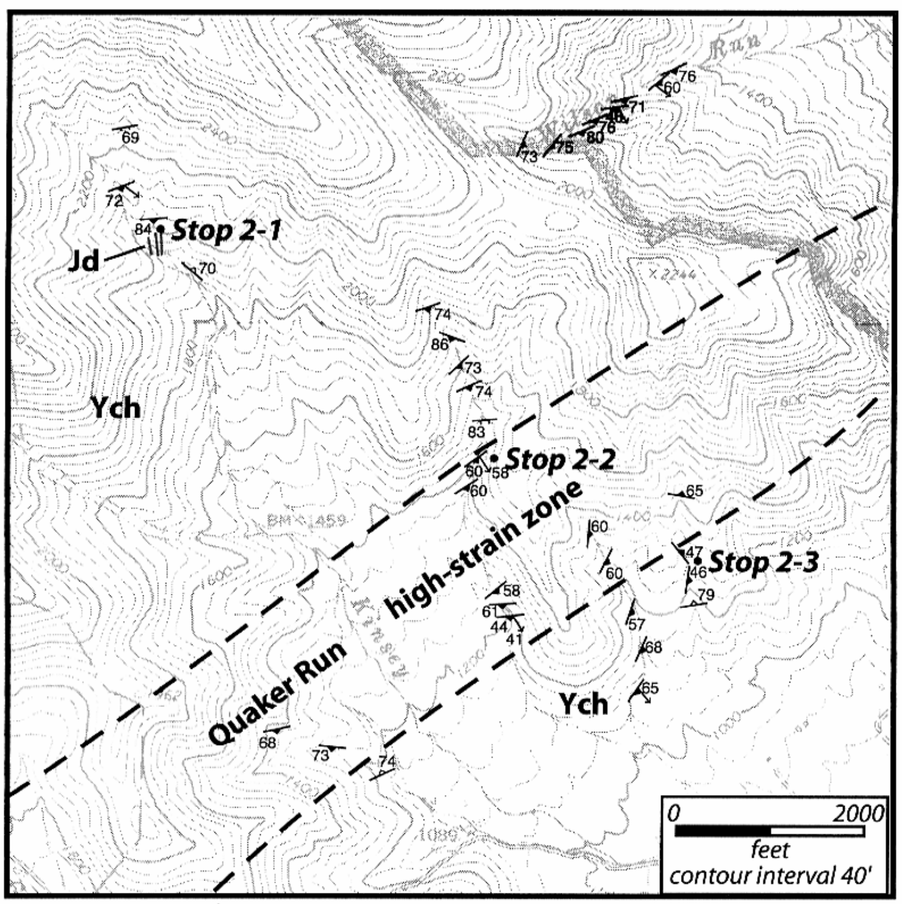


Figure 11. Topographic and geologic map of the eastern part of the Fletcher 7.5' quadrangle with Stops 2-1, 2, & 3. Ych- charnockite, Jd- diabase. Mapping by Gilmer, 1999.

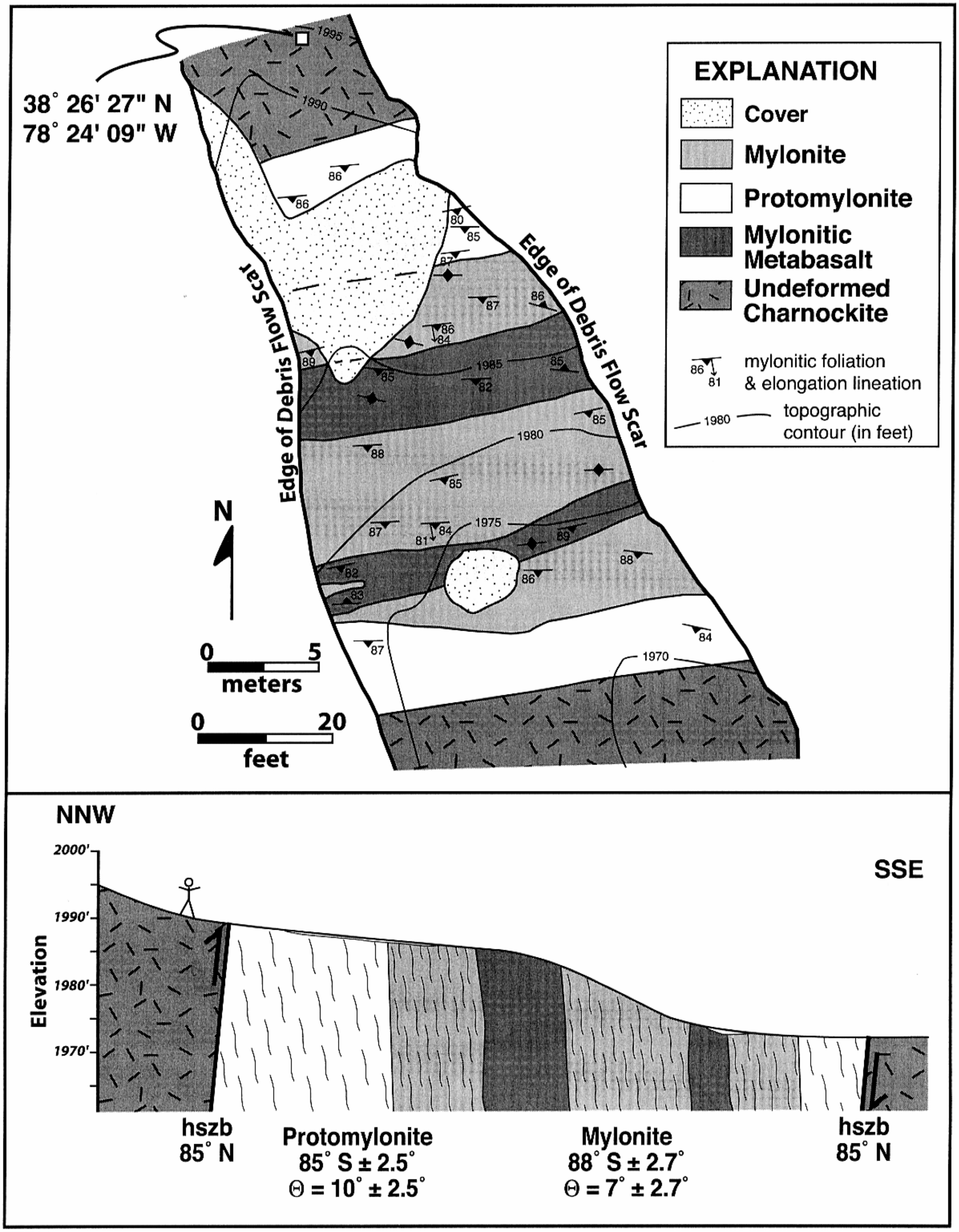


Figure 12. Geologic map and cross section of the Kinsey Run high-strain zone (from Bailey et al., *in review*).

The general lack of recrystallized quartz and the abundance of brittle microstructures in feldspar is consistent with greenschist facies deformation conditions (T ~ 300° - 400° C).

Strain was estimated from monocrystalline quartz lenses and ribbons using standard Rf/ϕ techniques with a hyperbolic stereonet (De Paor, 1988). 16-40 grains were measured per sample and strain in *XZ* sections averaged 3.8 (3.6- 4.0, n =2) for the protomylonites and 5.8 (5.5- 6.2, n =3) for the mylonites. Three-dimensional strains for all samples plot in the field of apparent flattening on a Flinn diagram (Fig. 13) and mylonitic samples consistently have lower K-values than protomylonites (K = 0.6 versus 0.8). In the *YZ* section (which approximates the outcrop surface) both metabasalt and epidotized leucocharnockite dikes exhibit pinch and swell structure or form boudins, consistent with true flattening strains and elongation parallel to *Y*. Undeformed charnockite, protomylonite, and mylonite are composed of approximately 61% SiO2 and have comparable amounts of other major elements. Based on the best-fit line (slope = 0.99) for the immobile elements (Al, Ti, and P) volume loss appears to be minimal (~1%, Fig. 13), although significant mineralogical changes occurred between the undeformed charnockite and mylonitic rocks.

Although there is a transition from undeformed charnockite to protomylonite, the Kinsey Run high-strain zone boundary dips steeply (~85°) to the northwest. Foliation orientations from the protomylonite and mylonite were averaged and yield Θ values of 10 ± 6° and 7 ± 6° respectively (Fig. 12). Protomylonite and mylonite samples have Wm values of 0.6 to 0.7 on the Rs/θ diagram, well into the field of general shear (Fig. 13). Minimum and maximum values for Wm, based on the uncertainty in the dip of the high-strain zone boundary and foliation, range from 0.5 to 0.85 (Fig. 13). Shear strain ranges from 1.5 to 1.8 in the protomylonite and 2.3 to 2.7 in the mylonite (Fig. 13). Although no independent estimate of Rs or γ was made for the metabasalt dikes, a maximum value was estimated using the contact relations with the surrounding mylonite. The pinch and swell nature of the metabasalt dikes suggests they were more competent than the surrounding mylonite, thus γ must be less than 2.3 for the metabasalt. Integrating shear strains across the Kinsey Run high-strain zone yields a total displacement of 60 ± 10 m. These results indicate that the Kinsey Run high-strain zone experienced a weak triclinic flow characterized by isochemical behavior and flattening strains that developed under general shear conditions.

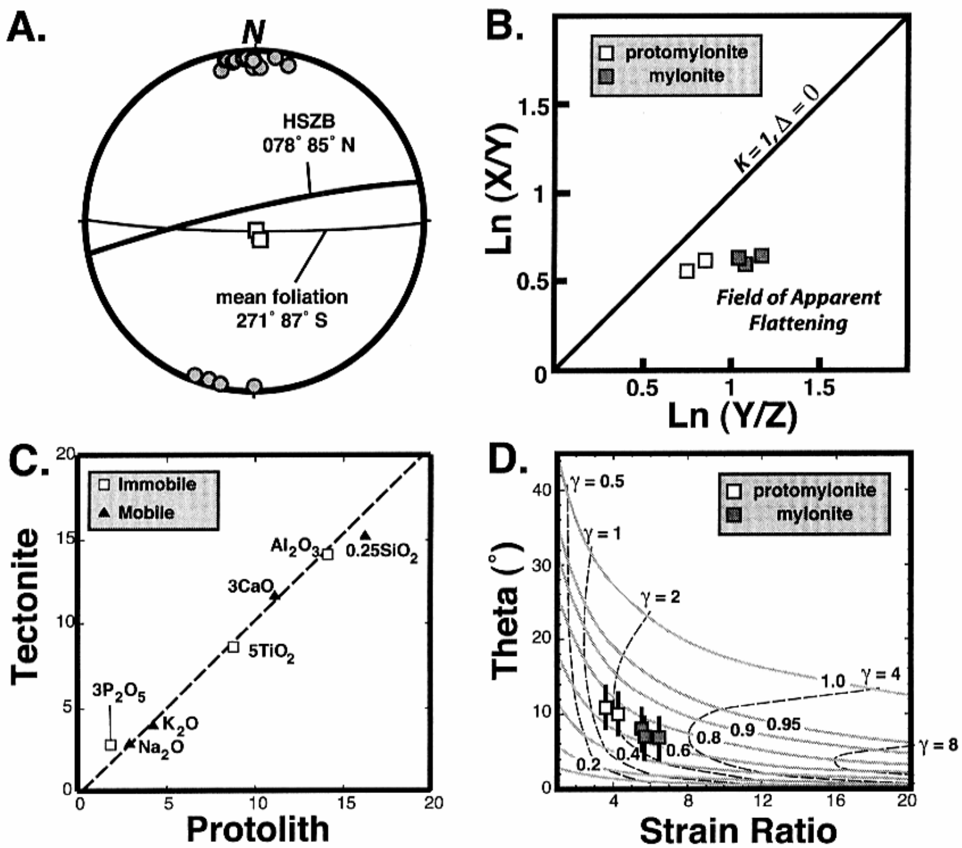


Figure 13. Equal-area stereogram of fabric elements of the upper Kinsey Run high-strain zone. Circles- poles to foliation, Open boxes- elongation lineation. Note the triclinic symmetry. **B.** Logarthimic Flinn diagram for quartz grain shapes. **C.** Isocon diagram. Dashed line represents best-fit line for immobile elements with a slope of 0.99 (r2 = 0.99), consistent with isochemical deformation. **D.** Rs/θ diagram. Wm = 0.65 (general shear) for protomylonites and mylonites. Thick black lines represent maximum uncertainties. From Bailey et. al., *in review.*

Return to the logging road and descend then cross to Stop 2-2

Stop 2-2
Quaker Run high-strain zone

This debris flow scar is one of the largest created by a storm in June 1995 and it exposes the base of the Quaker Run high-strain zone. Medium- to coarse-grained, massive to weakly foliated charnockite passes upward into protomylonite and mylonite. The contact between undeformed charnockite and mylonitic rocks strikes to the northeast and dips moderately to the southeast. The angle between the foliation and high-strain zone boundary is <10°. Mylonitic rocks are characterized by a southeast plunging downdip elongation lineation. Kinematic indicators, especially visible at this outcrop because of the vertical exposure, are plentiful and record top-to-the northwest sense of shear. There are, however, numerous back-rotated porphyroclasts (mostly feldspar megacrysts) suggesting that this zone experience general rather than simple shear. Leucopegmatite dikes are commonly oriented subparallel to foliation and exhibit pinch and swell structures along their margins. These competent dikes are cut by fibrous quartz filled extension fractures. A 2-4 m wide zone of fine-grained chlorite-bearing mylonite may be derived from a metabasalt.

The Quaker Run high-strain zone was defined by Berquist and Bailey (2000) for exposures a few kilometers to the north. At this latitude the Quaker Run high-strain zone is the thickest (0.5- 1.5 km) Paleozoic mylonite zone in the Blue Ridge and can be traced for over 30 km parallel to the regional trend. Strain is very heterogeneous. Based on estimates of shear strain (using values determined from other Blue Ridge mylonite zones) Berquist and Bailey estimated a total displacement of 1.5 ± 0.5 km across the Quaker Run zone.

Climb a rough trail over a low gap into the next debris flow channel to Stop 2-3

Stop 2-3
Variably deformed charnockite and metabasalt in the General's debris flow scar

The heterogenetity of the strain is evident in this debris flow channel as we pass from nearly massive charnockite to mylonite and narrow zones of ultramylonite and back into charnockite. Many of the narrow zones strike at high-angles to the regional trend (~035°), illustrating the anastomosing nature of these zones. Metabasalt dikes are also strongly deformed and in a few spots asymmetrically folded.

Upon reaching the base of the debris flow scar continue down the pasture (an ancient debris flow fan) past the cottage out to Rt. 670.

Stop 2-4
Granitoid and mylonite in the White Oak Run high-strain zone, Madison

At this old roadside quarry granitoids of the 705- 735 Ma Robertson River Igneous Suite have been transformed into mylonites (Fig. 14). Foliation, defined by aligned hornblende and quartz ribbons, strikes ~050° and dips steeply northwest. A downdip mineral elongation is well developed. Asymmetric structures record a down-to-the northwest (extensional) sense-of-shear. The White Oak Run high-strain zone is up to 1 km wide, ~ 15 km in length and occurs almost entirely within Robertson River granitoids. Feldspars in the mylonite have dynamically recrystallized. Biotite-garnet geothermometry, from mylonitic gneisses derived from basement units, indicates deformation temperatures of 500° C (Knight, 1999). In contrast to the abundant greenschist facies high-strain zones exposed further west in the Blue Ridge, the White Oak Run high-strain zone formed at the amphibolite facies. Knight and Bailey (1999) interpreted this zone to be a Neoproterozoic structure related to Iapetan rifting. Ar-Ar analysis of dynamically recrystallized hornblende yield messy spectra without a well-defined plateau and an integrated age of ~ 450 Ma.

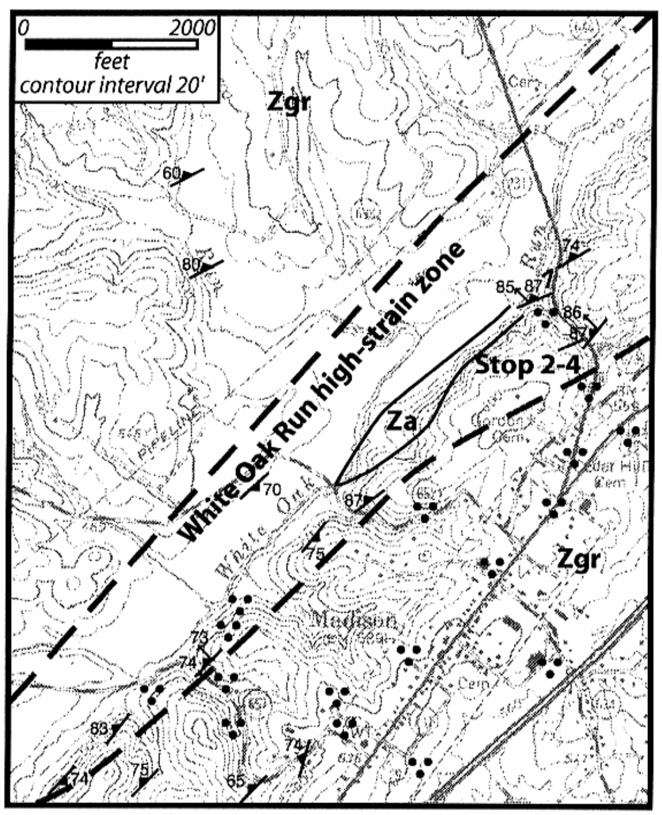


Figure 14. Topographic and geologic map of the southeastern part of the Madison 7.5' quadrangle with stop 2-4. Zgr- Robertson River granitoids, Za- amphibolite. Mapped by Knight, 1999.

Stop 2-5
Quaker Run high-strain zone 2,

Prior to the June 1995 this was a great exposure for the Blue Ridge, however the debris flow scars have changed our definition of great exposure in this region. Well foliated mylonitic gneiss derived from leucogranite and foliated metabasalt dikes are exposed at this outcrop. Kinematic indicators generally record top-to-the northwest movement, but the case can be made for general shear deformation.

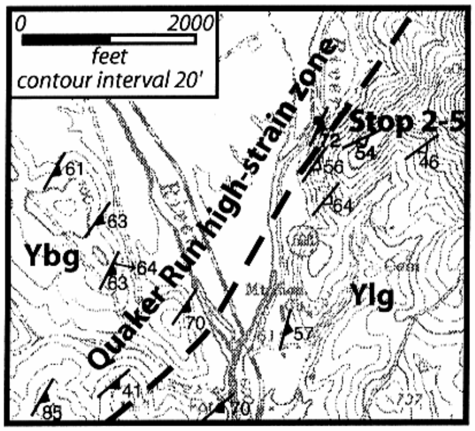


Figure 15. Topographic and geologic map of the central part of the Madison 7.5' quadrangle with stop 2-5. Ybg- biotite granitoid gneiss, Ylg- leucogranite and leucocratic gneiss. Mapped by Berquist, 2000.

Day 3

Stop 3-1

Grenvillian basement complex in debris flow scar Aylor Farm

This wonderful exposure illustrates cross-cutting relations amongst a number of Blue Ridge basement units (Fig. 16 and 17). The oldest rock at this outcrop is a leucocratic, medium-grained gneiss. The gneiss was intruded by coarse-grained leucogranite and leucogranite pegmatite. At the base of the outcrop a ~15 cm-thick dike of coarse-grained leucogranite is deformed into a series of tight folds with rounded hinges. The leucocratic gneiss has a weak foliation that strikes ~070°, dips steeply to the N, and is axial planar to the folded leucogranite dike. Simple line-length restoration of the dike indicates ~70% shortening in a NNW-SSE direction, but the fabric in the enclosing gneiss is barely visible. We interpret the medium-grained gneiss to have statically recrystallized at high temperatures after deformation, and thus it does not faithfully record the total strain history. This high-temperature foliation is well-developed in the leucocratic gneiss and is commonly overprinted by a foliation defined by aligned micas.

Further up this outcrop the leucocratic gneiss, leucogranite, and pegmatite are intruded by a 30-50 cm wide dike of medium- to fine-grained biotite granodiorite. This dike clearly post dates the deformation recorded in the gneiss and leucogranite. The mineralogy of the biotite granodiorite is similar to other basement units. The high-temperature fabric is interpreted to be Mesoproterozoic and associated with the Grenvillian event. Tollo and Aleinikoff (2001) obtained a U-Pb age of 1,150 Ma for a similar leucocratic gneiss approximately 8 km west of this exposure. The Old Rag granite, a coarse-grained leucogranite, exposed 15 km to the north in Shenandoah National Park has a U-Pb age of 1,065 ± 6 Ma (Tollo and Aleinikoff, 2001).

The basement units are cut by a northeast-striking ~5 m wide dike of porphyritic hornblende gabbro. The gabbro is composed of hornblende, extensively altered plagioclase, epidote, and chlorite. The dike is part of suite of mafic to ultramafic igneous rocks that intrude Blue Ridge basement and Neoproterozoic metasedimentary units.

A set of discrete high-strain zones cut all the basement units. These zones range from a few centimeters in thickness to mm-scale. Where discernible mineral elongation lineations plunge obliquely down-dip. The apparent offset, as illustrated on the sub-horizontal outcrop surface, is dextral, but clearly there was out-of-section movement. The mineralogy and microstructures in the high-strain zones are consistent with greenschist facies conditions and we interpret these zones to be Paleozoic.

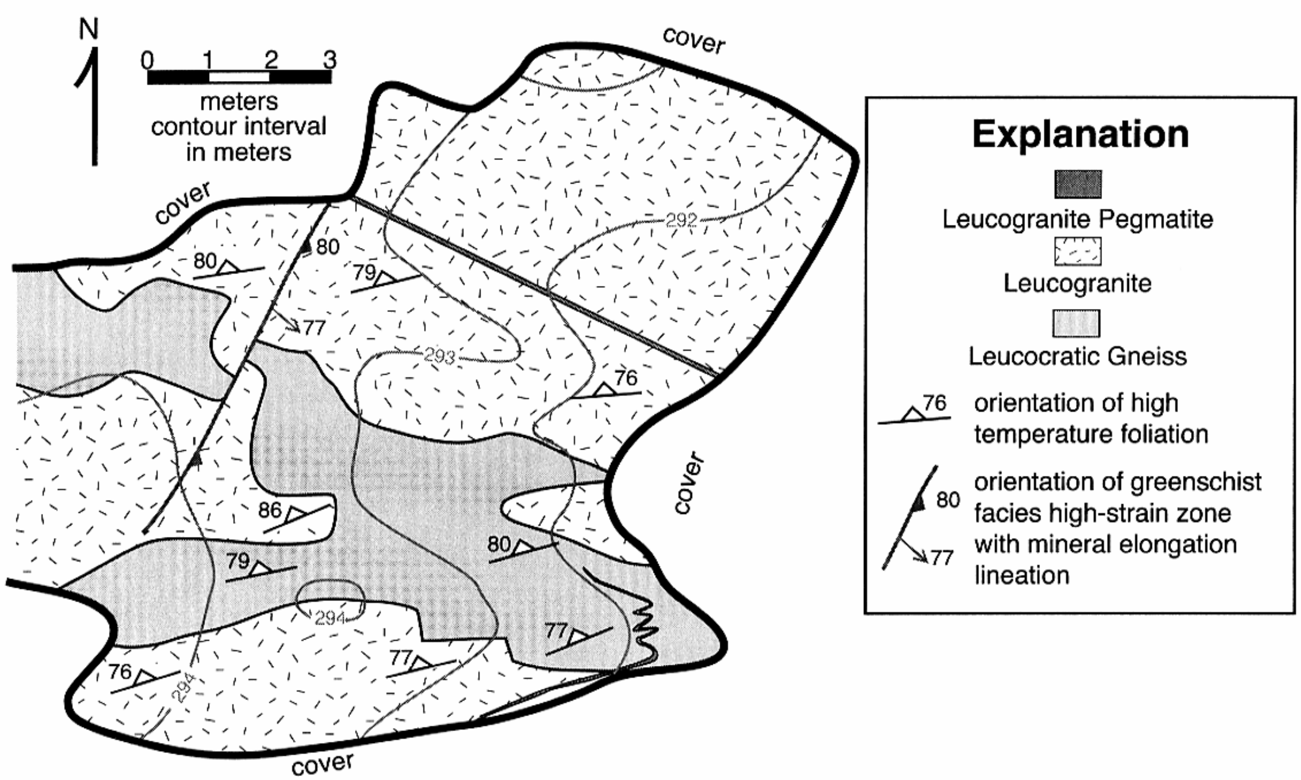


Figure16. Geologic map of the lower part of the Aylor Farm debris flow scar outcrop.

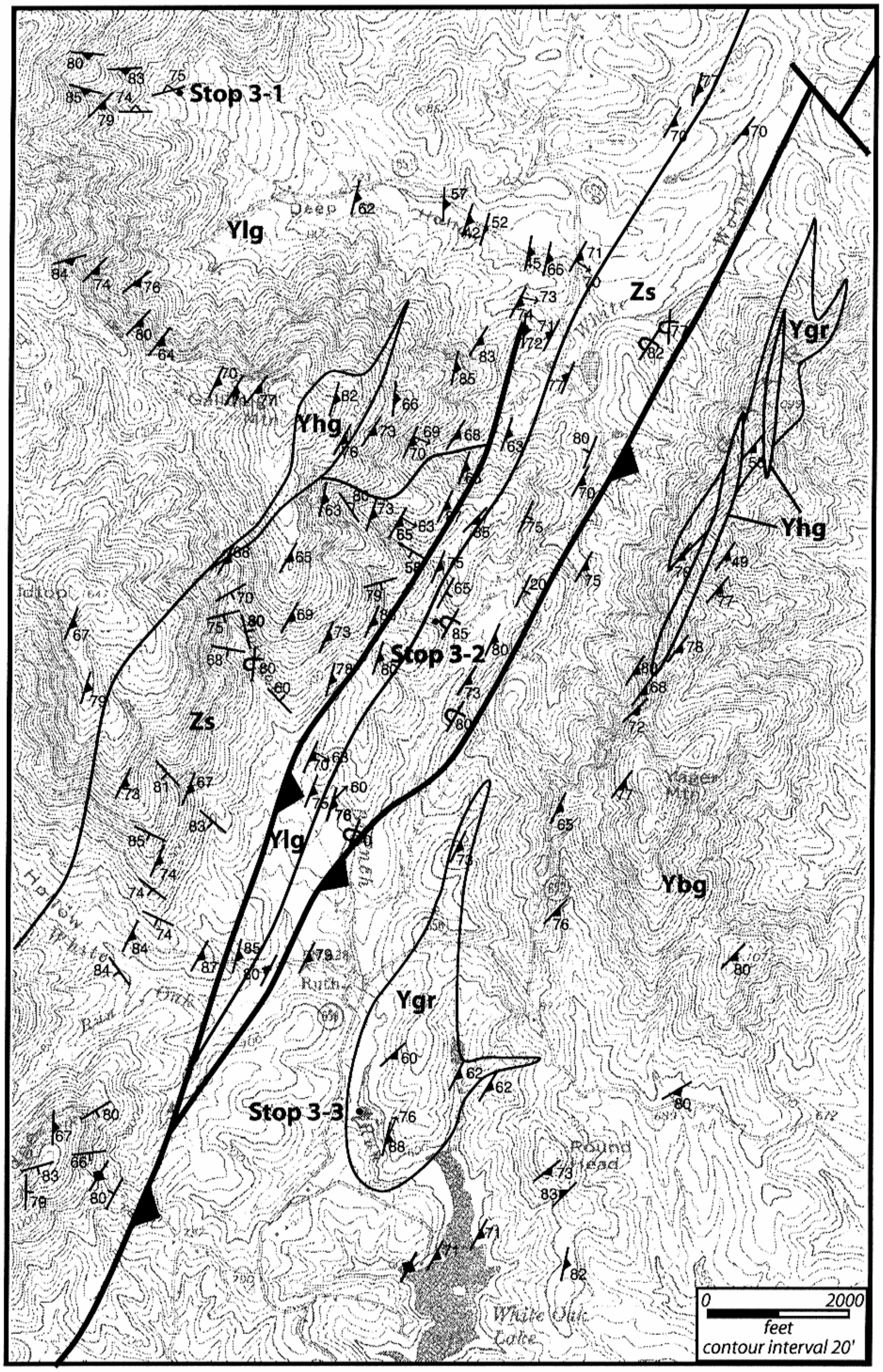


Figure 17. Topographic and geologic map of the central part of the Madison 7.5' quadrangle. Zhg- hornblende metagabbro, Zs- Mechum River Formation, Ygr- Coarse-grained alkali feldspar granite, Ybg- biotite granitoid gneiss, Ylg- leucocratic gneiss and leucogranite.

Stop 3-2

Arkosic wackes, Neoproterozoic Mechum River Formation

The Neoproterozoic Mechum River Formation is exposed in a narrow (0.5- 2 km) wide NE-SW trending belt that is surrounded by Grenvillian basement. At the northwestern contact of this belt metasedimentary rocks unconformably overlie the basement contact. At the southeastern contact basement rocks structurally overlie the Mechum River Formation along a reverse fault.

At this exposure granular arkosic wackes of the Mechum River Formation crop out in repeating, graded packages 20-50 cm thick and are commonly overlain by thinner packages of fine-grained arkosic wackes with a well-developed foliation defined by aligned micas. The penetrative foliation is axial planar to map-scale folds. At this location bedding strikes to the NE, dips steeply to the SE, and is overturned. The clast assemblage includes perthitic feldspar, quartz, sphene, and ilmenite. The matrix is dominated by fine-grained quartz, muscovite, biotite, and epidote. The mineral assemblage and microstructures preserved in these rocks are consistent with lower to middle greenschist facies conditions.

Shotwell and Bailey (2000) measured strain in 36 samples of Mechum River arkose; whole rock strain using both quartz and feldspar clasts ranged from 1.1- 2.4 in *XZ* sections. In most samples quartz records 10-30% higher strain than feldspar. Cross section restoration indicates ~30% shortening due to map-scale folding, while grain scale processes accomplished up to an additional ~30% shortening.

Stop 3-3

Yowell Farm, narrow high-strain zones cutting coarse-grained alkali feldspar granite,

A narrow (<30 cm) high-strain zone cuts coarse-grained alkali feldspar granite in a pavement exposure on the Yowell Farm (Fig. 17). The main zone of mylonite strikes north-northeast, dips steeply to the southeast and thins to the northeast (Fig. 18). Away from the main high-strain zone the granite is cut by a series of anastomosing mm- to cm-scale high-strain zones. The strike of the mylonitic foliation is subparallel to the high-strain zone boundary and generally dips less steeply to the east (Fig. 18). A down-dip mineral elongation lineation is present and the overall symmetry of the fabric elements is monoclinic (Fig. 18). Asymmetric structures are common on sections normal to the foliation and parallel the lineation (XZ) and symmetric structures predominant on sections normal to both the foliation and lineation (YZ).

Leucogranitic country rock is composed predominantly of 0.5 to 2 cm diameter alkali feldspar (~50%) and quartz (~30%) with minor plagioclase, muscovite, biotite, and epidote. Away from the high-strain zone the leucogranite is massive to weakly foliated (defined by aligned quartz). The transition to mylonitic rocks in the high-strain zone is sharp and the foliation defined by aligned muscovite, lenses and ribbons of quartz, and elongate fractured alkali feldspars. Muscovite and quartz form ~75% of the mylonite with alkali feldspar comprising <20%. In thin section, feldspars are extensively fractured, whereas quartz forms lenses and ribbons with a strong crystallographic preferred orientation and well-developed core and mantle textures. Although some quartz has recrystallized, lenses and ribbons are distinct from the fine-grained matrix. Feldspar and quartz microstructures are consistent with mid-greenschist facies (T ~400° C) conditions.

Sectional strain (Rs) was estimated from quartz grain shapes using the R/ϕ method with a hyperbolic stereonet (De Paor, 1988). Fabrics in the leucogranite outside the high-strain zone were only very weakly developed (Rs = 1.1-1.3) indicating that the rock had, at most, a very weak fabric prior to mylonitization. Quartz grain shapes were measured for three individual samples in the high-strain zone (including two 2 cm-diameter drill cores). Individual samples had too few measurable grains and data was combined to form a single composite Rs value. The Rs value for quartz lenses and ribbons in the *XZ* section is 13 and 4 in the *YZ* section. Combined aspect ratios from the *XZ* and *YZ* sections, yield a *XY* ratio of 3.2, consistent with a modest flattening strain (K = 0.8).

The finite (or mean) kinematic vorticity number (Wm of Passchier, 1986) was estimated based on the strain ratio (Rs) and the angle between the high-strain zone boundary and the long axis of the strain ellipsoid (θ) (Fossen and Tikoff, 1993; Bailey et al., 1999). The high-strain zone boundary dips 77° ± 4° to the southeast and the mean foliation dips 72° forming a θ angle of 5°. Quartz grain shape and foliation orientation data yield a Wm value of 0.8 (general shear) with a shear strain of ~6 (Fig. 19). Maximum uncertainties for Wm, based on the standard deviation of quartz grain shape aspect ratios and differences in the high-strain zone boundary and foliation dip angles, ranged from 0.1 to 0.95 (Fig. 19). Even at the limits of uncertainty the Yowell Farm high-strain zone still plots well within the field of general shear on the Rs/θ diagram (Fig. 19).

The undeformed leucogranite and mylonite are composed of 71-72% SiO2 and have similar amounts of other major elements (Fig. 19). Bulk-rock volume changes (D) associated with deformation were assessed by comparing immobile elements (Al, Ti, and P) in the protolith and tectonite. Based on the best-fit line (slope = 1.014, r = 0.98) for the immobile elements volume loss appears to be minimal (~1%, Fig. 19). However, the mineralogical differences from protolith to mylonite are consistent with alkali feldspar being replaced by muscovite and quartz. Collectively, fabric and chemical data from the narrow Yowell Farm high-strain zone suggest bulk isochemical (isovolumetric?) behavior, modest flattening strains, and a general shear deformation with a monoclinic symmetry.

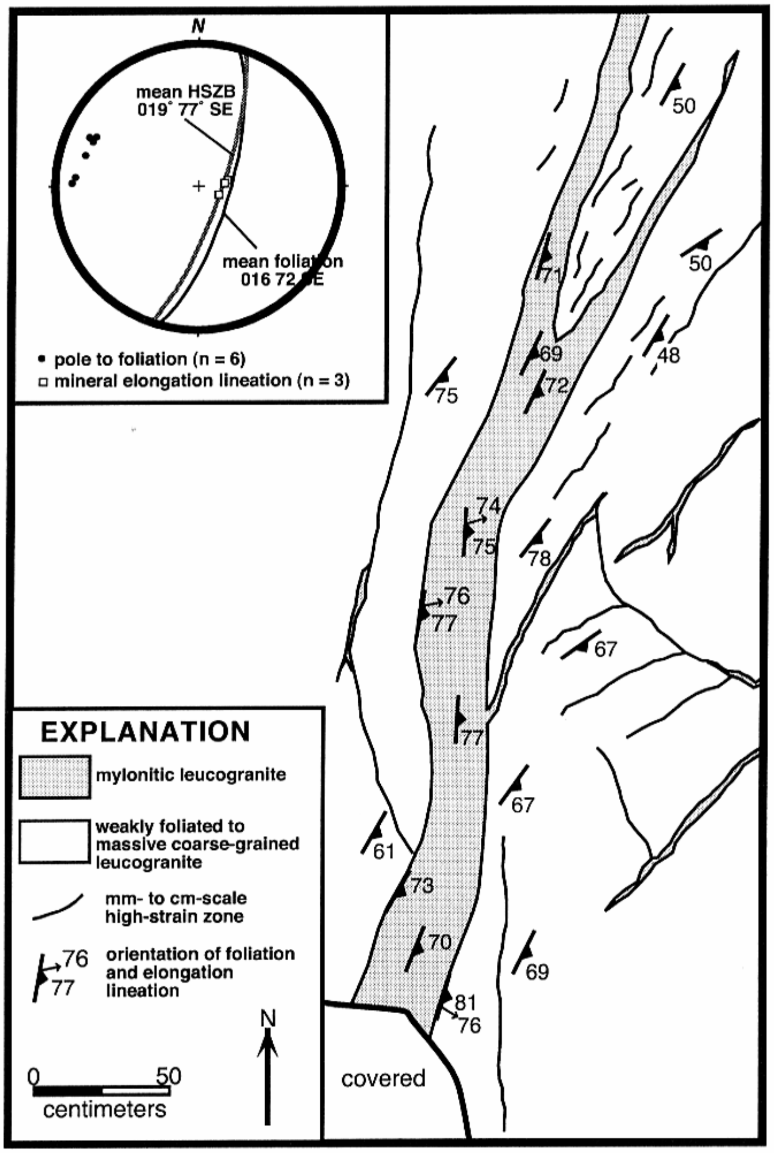


Figure 18. Geologic map of the Yowell Farm high-strain zone and equal-area stereogram of fabric elements. Outcrop surface is a subhorizontal pavement exposure from Bailey et al., *in review.*

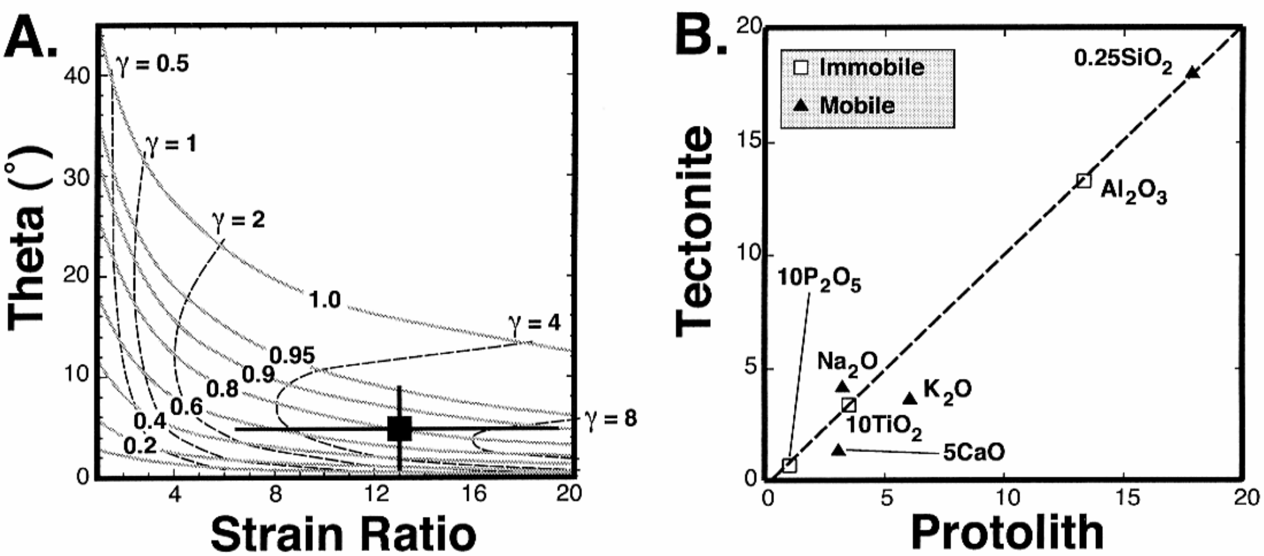


Figure 19. Rs/θ diagram for the Yowell Farm high-strain zone. Wm = 0.8 (general shear), γ ~6. Thick black lines represent maximum uncertainties: based on standard deviation of quartz grain shapes and total range of foliation dip angles. B. Isocon diagram for the Yowell Farm high-strain zone. Dashed line represents best-fit line for immobile elements with a slope of 1.014 (r2 = 0.98), consistent with isochemical deformation. from Bailey et al., *in review.*

Stop 3-4

Garth Run high-strain zone

The Garth Run high-strain zone is well exposed in a 200 m-long outcrop that was scoured out along the channel of Garth Run during the June 1995 storm (Fig. 20). The zone strikes north-northwest to north-northeast and joins the 1-2 km wide Quaker Run high-strain zone to the north and tips out to the south (Berquist and Bailey, 2000). Although the high-strain zone boundary is not exposed, outcrop data indicate the zone is approximately 125 m thick (Fig. 20). To the west the zone is bounded by medium- to coarse-grained, equigranular to porphyritic charnockite. A medium-grained layered gneiss is the dominant rock type east of the Garth Run high-strain zone and is intruded by a series of fine- to coarse-grained leucogranite dikes ranging from 0.3 to 5 m in thickness.

Rocks exposed in the Garth Run high-strain zone are dominantly porphyroclast-bearing protomylonites and mylonites derived from charnockite and alkali feldspar granite with subordinate amounts of equigranular mylonites derived from gneiss. Finely-layered mylonitic leucogneiss, leucogranite, and well-foliated fine-grained metabasalt occur as tabular to lenticular bodies (0.2-2 m thick) throughout the high-strain zone. These tabular bodies are both subparallel to and slightly discordant to the foliation. Foliation, defined by mica-rich surfaces, elongate quartz grains, and fractured feldspars, strikes 345° to 010° and dips between 30-50° to the east (Fig. 21). A mineral elongation lineation plunges down-dip or obliquely to the east-northeast (Fig. 21), however mineral elongation lineations are not present everywhere.

Asymmetric structures such as σ and δ porphyroclasts, shear bands, and asymmetric boudins are common in the Garth Run high-strain zone. Sections normal to foliation and parallel to lineation consistently preserve structures with a top-to-the west asymmetry (reverse sense of shear). Sections normal to both foliation and lineation also preserve asymmetric structures indicating a sinistral (when viewed looking down the elongation lineation), top-to-the north sense of shear (strike-parallel). Asymmetric structures, both parallel and perpendicular to the mineral elongation lineation, are consistent with a triclinic deformation symmetry for the Garth Run high-strain zone.

Sheath folds occur in the finely-layered mylonitic leucogneiss and sheath axes plunge moderately east, parallel to the mineral elongation lineation. In exposures normal to foliation and parallel to lineation sheath folds are elliptical. Coarse-grained leucogranitic bodies display pinch and swell structures and are commonly isolated as lozenge-shaped boudins surrounded by porphyroclastic mylonites. Leucogranitic boudins record elongation both parallel to and normal to the east-northeast plunging elongation lineation indicating bulk extension in Y and X. Leucogranites have a weak foliation and are cut by numerous transgranular fractures. At two locations slightly discordant tabular boudinaged leucogranitic dikes are folded. At a few locations the mylonitic foliation is kinked in narrow (2-5 cm wide) bands.

Feldspar microstructures in the Garth Run high-strain zone include intra- and inter- granular fractures and grain boundary mantles of muscovite and quartz. Quartz forms both monocrystalline and polycrystalline ribbons with a moderate to strong XPO. In the mylonitic leucogneiss, quartz is completely recrystallized forming a fine-grained matrix between alkali feldspar grains that display no evidence of crystal plasticity. As with the other high-strain zones examined in this study, microstructures from the Garth Run zone are consistent with greenschist facies (~400° C) deformation.

Strain was estimated from discrete quartz lenses and ribbons in mylonitic rocks derived from charnockite. Rs values range from 5 to 23 in *XZ* sections and 3 to 18 in *YZ* sections. Three-dimensional strains for all samples plot in the field of apparent flattening on a Flinn diagram with K-values between 0.05 -0.45 (Fig. 22). Although no chemical data was collected for the Garth Run samples, thin sections of Garth Run charnockite and mylonite are similar to Kinsey Run samples that experienced isochemical behavior. Even with volume losses of 60% (the maximum reported by Bailey et al., 1994 for Blue Ridge high-strain zones) most Garth Run mylonites record true flattening strains (Fig. 22).

As the high-strain zone boundary was not exposed, Wm could not be estimated using the Rs/θ method. Rather Wm-values were obtained with the porphyroclast hyperbolic distribution (PHD) method of Simpson and De Paor (1993) on ultramylonite thin sections and joint faces (normal to foliation and parallel to the mineral elongation lineation) with well-exposed porphyroclasts. During general shear deformation some porphyroclasts forward-rotate while others backward-rotate. The orientation and aspect ratio of forward- and backward-rotated grains are plotted on a hyperbolic net (De Paor, 1988) and the hyperbola that separates back rotated grains from forward rotated grains constructed. Wm is the cosine of the interlimb angle of the hyperbola, where each limb represents a flow apophysis or eigenvector (Bobyarchick, 1986; Simpson and De Paor, 1993). The smallest hyperbola separating the fields of backward-rotated grains from forward-rotated grains range from 55° and 68° (n =4), yielding Wm-values of 0.6 to 0.4, indicating general shear deformation (Fig 22). Shear strain is conservatively estimated at 4 (based on an Rs ~10 and Wm ~ 0.5) and integrates to a total displacement of ~500 m for the Garth Run high-strain zone, however discrete brittle faults with displacements of 0.2 to 2 m cut the mylonitic fabrics.

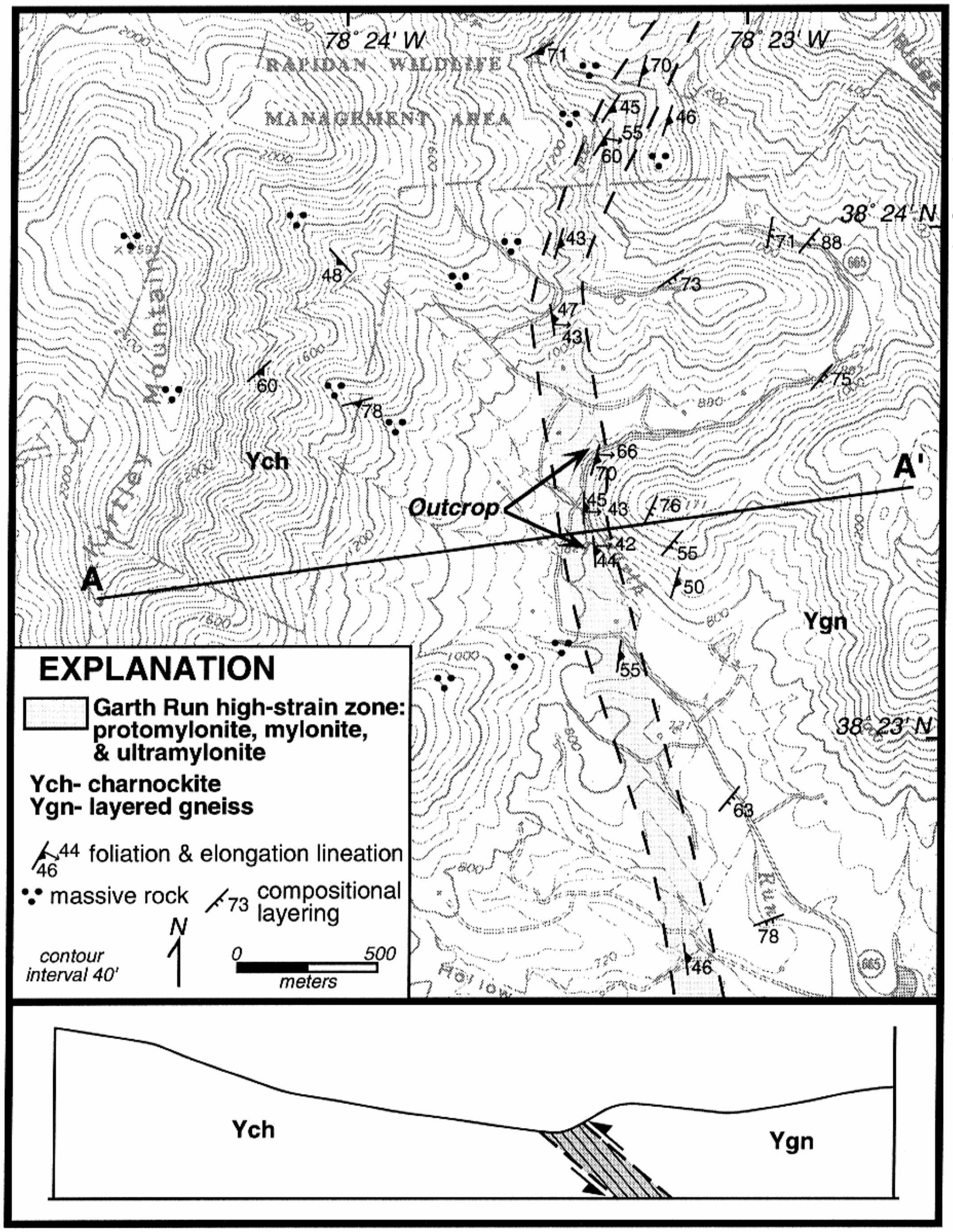


Figure 20. Topographic and geologic map of the eastern part of the Fletcher 7.5' quadrangle with Stop 3-4

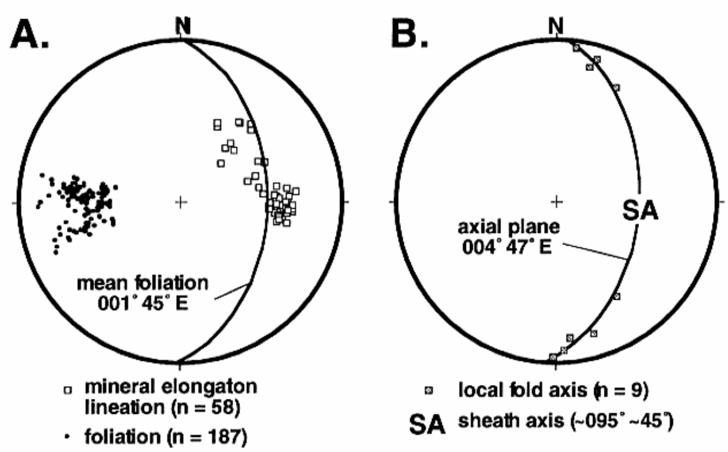
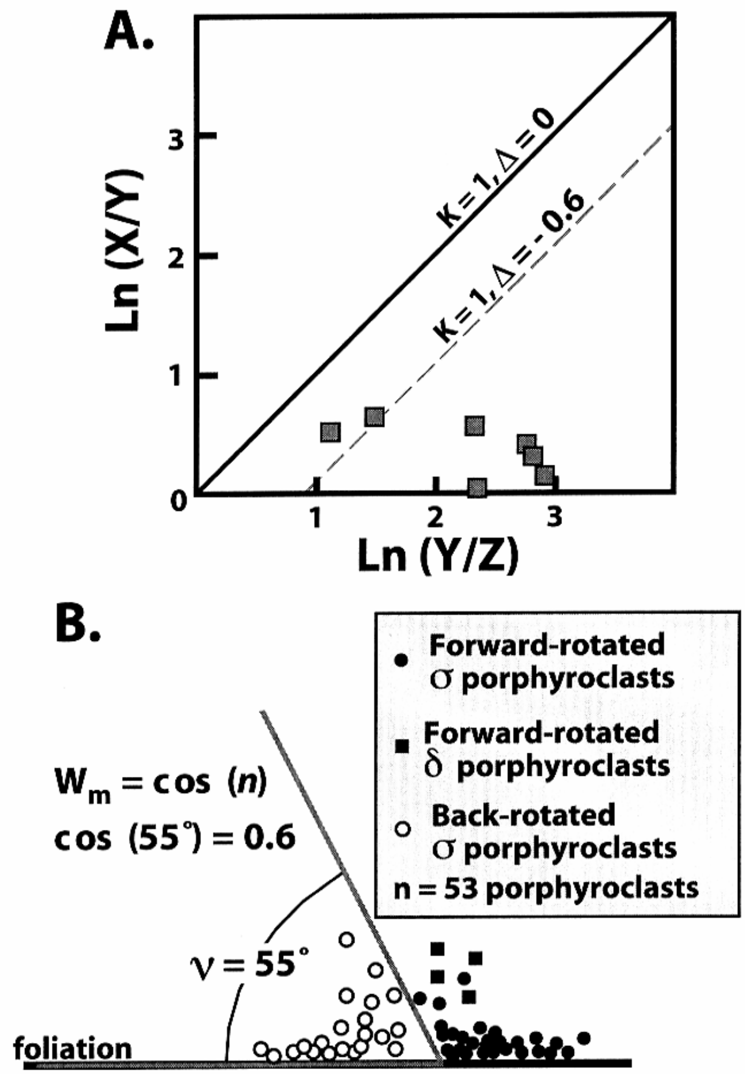


Figure 21. Stereogram of fabric elements for the Garth Run high-strain zone

Folded leucogranite boudins may offer a clue about the progressive deformation history at the Garth Run high-strain zone. These structures formed when competent leucogranite dikes were first elongated and then shortened. During a progressive, steady-state deformation (regardless of whether it is simple shear, general shear, or pure shear) material rotates from the field of shortening into the field of extension, a progression that will not fold boudins. Folded boudins are generally interpreted to develop by polyphase deformation, such that material elongated during the first deformation is shortened by a second deformation with a different orientation. However, a change in the incremental vorticity will cause some material that was originally deformed in the field of extension to move into the field of shortening. There are no cross cutting ductile fabrics in the Garth Run high-strain zone and no compelling evidence for two episodes of deformation. We favor a model in which the Garth Run high-strain zone undergoes a single deformation that changes from simple shear dominated to a pure shear dominated with time (Fig. 22), but the finite vorticity records bulk general shear (Wm~ 0.5).



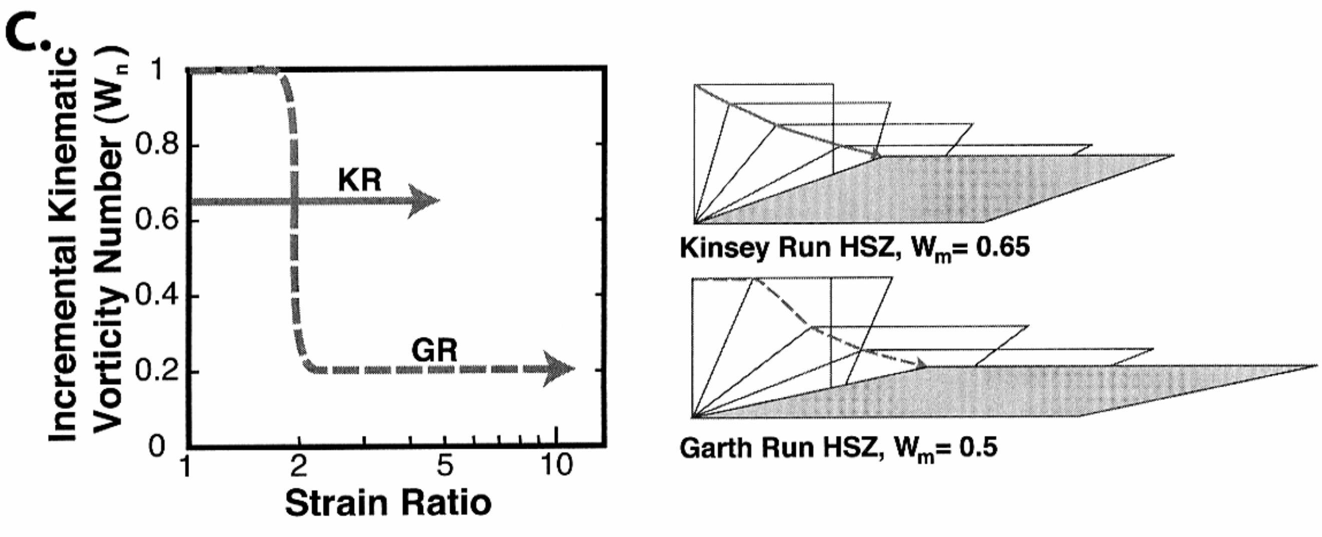


Figure 22A. Flinn diagram for quartz fabric from the Garth Run high-strain zone.
B. Hyperbolic stereonet of porphyroclast distribution.
C. Generalized strain path for Kinsey Run and Grath Run high-strain zones.
from Bailey et. al., *in review*.

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