Dickinson College

Department of Earth Sciences

A STUDY OF PROGRESSIVE DEFORMATION IN THE HINGE OF THE PENNSYLVANIA SALIENT; SOUTHERN VALLEY AND RIDGE PROVINCE, PERRY COUNTY, PA

A Thesis in

Geology

By

Marci Allison Wills

Class of 2010

Submitted in Partial Fulfillment

of the Requirements

for the Degree of

Bachelor of Science with Honors

May 2010

ABSTRACT

This study characterizes the deformational history of a ~700 m long exposure in the Upper Devonian Irish Springs member of the Catskill Formation in the southern Valley and Ridge province. The study area is located along the west shore of the Susquehanna River in the hinge of the Pennsylvania salient. The sandstones, siltstones, and shales contain 3rd and 4th order folds, conjugate wedge faults, thrust faults, joints, cleavage, and grain-scale finite strain indicators. This suite of structures records a complex deformational history similar to the Bear Valley sequence of progressive deformation (Nickelsen, 1979) observed 40 km to the ENE. Analysis of the orientations of structural features from successive stages of the progressive deformation is used to constrain the orientations of the maximum shortening direction (MSD) and to distinguish between a single-stage (Gray and Stamatakos, 1997) and two-stage (Wise, 2004) model for the development of the Pennsylvania salient. Consistently oriented MSDs ranging from ~335° to 351° at the study area parallel findings 100 km to the NW in the hinge of the salient at the Appalachian structural front (Spiker and Gray, 1997). This is distinctly different from the CW rotations of MSDs seen in the Reading Prong (e.g. Gray and Mitra, 1993) and CCW rotation of MSDs in the Blue Ridge (e.g., Nickelsen, 2009), suggesting that the hinge of the Pennsylvania salient coincides with an axis of no rotation of MSDs, consistent with the single-stage model (Gray and Stamatakos, 1997).

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

ACKNOWLEDGEMENTS

First, I would like to thank my advisor, Pete Sak, for his invaluable help in taking on this project. Pete's advice, encouragement, patience, high expectations and early deadlines were vital in its successful completion. In addition, I thank him for his field assistance and skillful weedwhacking, as well for the use of his office and four computer screens. Thanks also to Gwen Dunnington, Mary-Beth Gray, Marcus Key, and Nate Lorentz for their helpful comments and revisions. Thanks to Rob Dean who taught me how to make thin sections and to Jake Davidson, Gwen Dunnington and Chrissy Miller who supplied me with their field data from the fall of 2009, some of which I used in a portion of this project. I would also like to thank each of my geology professors, Ben Edwards, Ewan Fordyce, Marcus Key, Jeff Niemitz, Meaghan Pollock, Pete Sak, and Virginia Toy for showing me how awesome geology is. Finally, thanks to my parents, John and Becky Wills, and also to all my fellow Dickinson geology majors for their support and so many good memories.

INTRODUCTION

The Pennsylvania salient of the central Appalachian Mountains is one of the most widely cited examples of orogenic curvature, a feature common to many mountain belts worldwide (Macedo and Marshak, 1999; Weil and Sussman, 2004; Yonkee and Weil, 2009). Curves in mountain belts are classified as salients or recesses depending on their orientation with respect to the tectonic transport direction; salients are oriented concave towards the hinterland and recesses concave towards the foreland (Thomas, 1983). The Pennsylvania salient is concave eastward and spans approximately 300 km across south central Pennsylvania and Maryland (Figure 1). It is defined by three roughly linear segments of different trend: the northern Reading Prong (065°-085°), the central segment (055°-060°), and the southern Blue Ridge (020°-085°) (Gray and Stamatakos, 1997). Despite over 150 years of study (Rogers, 1858), the mechanisms behind the formation of the Pennsylvania salient remain enigmatic.

The origins of orogenic curvature are not well understood although numerous factors affecting the geometry of mountain belts have been investigated. These include an irregular shape of the pre-existing continental margin (Thomas, 1983), variations in the thickness of the predeformational sedimentary basin (Thomas, 1977; Hatcher, 1989; Marshak and Wilkerson, 1992; Lawton et al., 1994; Boyer, 1995), strength of the detachment surface (e.g., Davis and Engelder, 1985), the presence of hinterland indentors (Pavoni, 1986; Laubscher, 1972; Tapponnier and Molnar, 1976; Gibb, 1978; 1983), strike-slip faulting across the main trend of the fold-thrust belt, and multiple non-parallel episodes of deformation (Marcedo and Marshak, 1999; Marshak 2004).

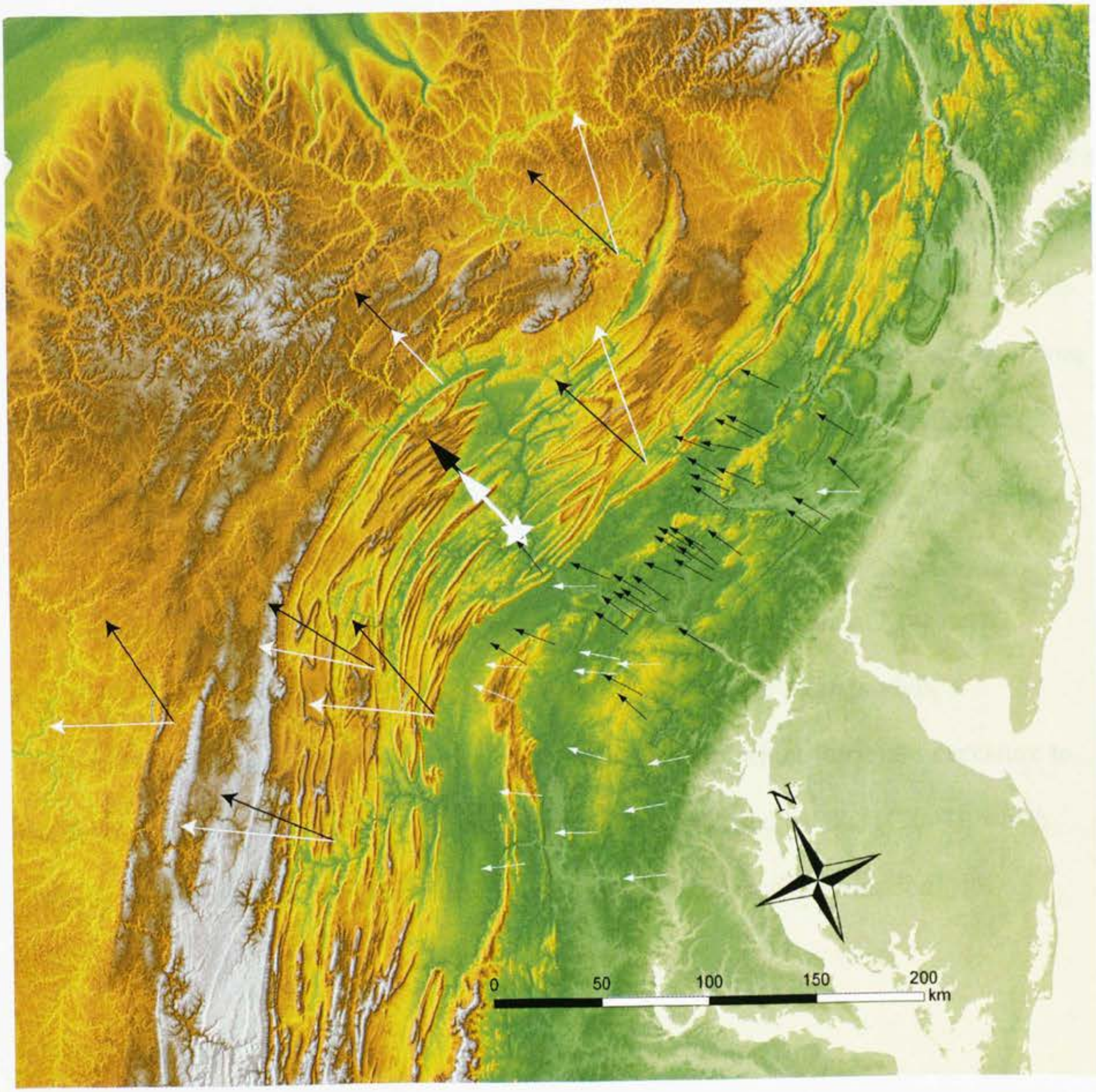


Figure 1: Digital elevation map of the central Appalachian region with arrows showing maximum shortening directions (MSDs) across the Pennsylvania salient. Shorter arrows in the Piedmont province (compiled by Wise and Werner, 2004) are divided into two azimuthal groups corresponding to early (black) and late (white) stages of deformation. Longer arrows in the Valley and Ridge and Appalachian Plateau provinces (compiled by Gray and Stamatakos, 1997) indicate progressive rotation of MSD with time from the earliest (black) to latest (white) stages of deformation. The white star and bold set of arrows shows the MSDs determined at the site of this study. Base image modified after Miller et al, 2009.

Additionally, the timing of curvature relative to orogenesis is critical to understanding the origin of any salient or recess. Orogenic curves are generally assigned to one of three categories: primary, secondary and progressive arcs (Marshak, 2004). Those that develop in a curved shape from the onset of deformation are called primary, or non-rotational, arcs. Secondary, or rotational, arcs are initially linear mountain belts that have been subsequently bent by another force. Primary and secondary arcs represent two end member models. Those curves that fall in between acquire curvature continuously throughout the evolution of the mountain belt and are classified as progressive arcs (Marshak, 2004; Weil and Sussman, 2004).

Numerous models for the development of the Pennsylvania salient have been suggested. This study will attempt to differentiate between two of these models; a progressive model, proposed by Gray and Stamatakos (1997), which attributes curvature to variation in sediment thickness, and a secondary model, proposed by Wise (2004), which explains the curvature on the basis of two non-parallel episodes of deformation.

GEOLOGIC SETTING

The Appalachian Mountains extend for > 3000 km along the east coast of North America from Newfoundland to Alabama and are a prime example of a non-linear mountain belt. The over one-billion-year tectonic history of the Appalachians includes five compressional orogenies and two extensional episodes (Faill, 1997a; 1997b; 1998). The first of these, the Grenville Orogeny, built the supercontinent Rodina during the middle Proterozoic. Next, an episode of extensive rifting in the late Proterozoic to early Cambrian broke apart Rodina into the continents of Laurentia and Gondwana (Faill, 1997a). Rifting was followed by the Potomac, Taconic, and Acadian orogenies in which a subduction zone accreted island arcs onto the eastern edge Laurentia during the Cambrian to Early Devonian (Faill, 1997b). In the central Appalachians, mountain building culminated in the early Permian Alleghany Orogeny which marks the closing of the Iapatus Ocean and suturing of Gondwana and Laurentia to form the Pangean supercontinent (Faill, 1998). Finally, an episode of rifting beginning in the late Triassic to early Jurassic broke apart Pangea with the opening of the Atlantic Ocean. The resulting Appalachian mountain range of eastern North America consists of alternating salients and recesses along its length. From north to south, the Newfoundland, Quebec, Pennsylvania and Tennessee salients are spaced approximately 400 to 500 km apart with the St. Lawrence, New York, Virginia and Alabama recesses spanning between them (Thomas 1983).

The Pennsylvania salient is one of the more prominent curves of the Appalachians in map view. The region surrounding the salient is comprised of three physiographic provinces differentiated by rock age and topography (Figure 2). Hinterland to the mountain belt, the Piedmont province of southeastern Pennsylvania is comprised of rocks Ordovician age and older (Figure 2B; U.S.G.S., 2009). The salient is most recognizable in map view in the Valley and Ridge province where shortening in the Ordovician to Pennsylvanian rocks created folds with 50 - 60 km long hinges and wavelengths of ~7-12 km. These folds are characterized by narrow hinges relative to their wavelength. Differential erosion of a variety of sedimentary rock types of the Valley and Ridge province (limestones, shales, siltstones, sandstones and conglomerates) has resulted in a first order topography of alternating parallel ridges and valleys which define the arc of the Pennsylvania salient across south central Pennsylvania. In the Susquehanna River valley in the center of the salient, the ridges trend

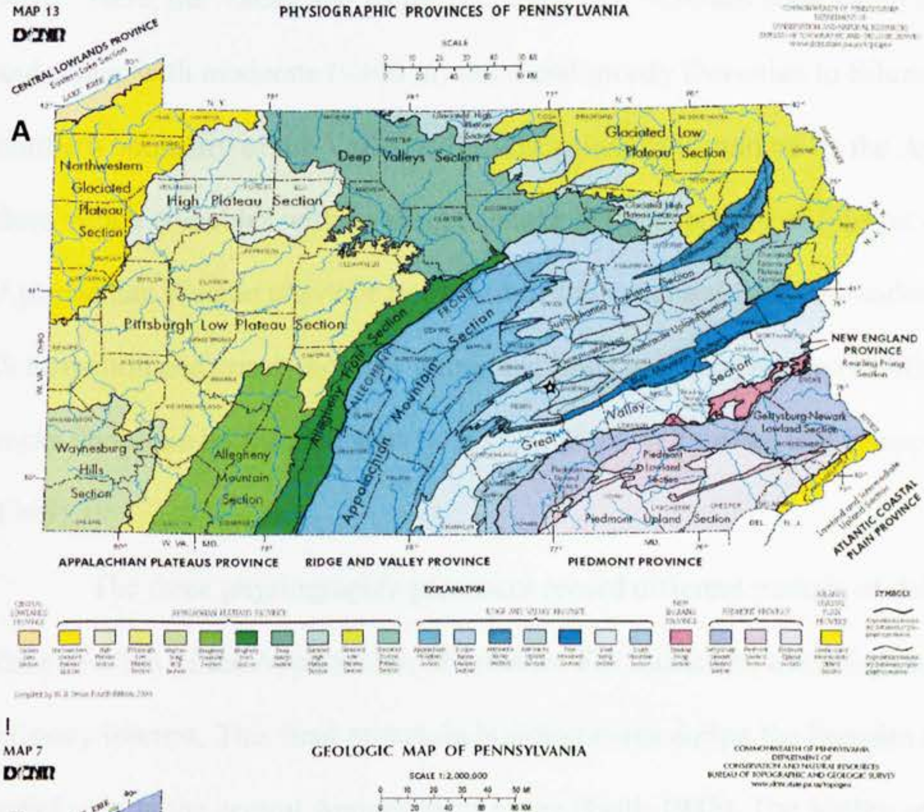


Figure 2: A: Geologic Map of Pennsylvania showing the physiographic provinces. Green- Appalachian Plateau province; Blue- Valley and Ridge province; Purple- Piedmont province; Star- study area. **B:** Geologic map of Pennsylvania showing the distribution of rock ages.(Pennsylvania Geology Survey, 2007; U.S.G.S., 2009).

~070°. Here, the Valley and Ridge is defined as a ~110 km wide swath of alternating valleys and ridges with moderate (<400 m) relief and mostly Devonian to Silurian age rocks. The northern boundary of the Valley and Ridge province is marked by the Appalachian structural front which coincides with an abrupt change in the geometry of folds at the surface. The Appalachian Plateau province northwest of the structural front has undergone less deformation and consists of low amplitude folds in Pennsylvanian to Mississippian to age rocks underlain by Silurian salts (U.S.G.S., 2009; Willtschko and Chapple, 1977; Rodgers, 1949).

The three physiographic provinces record different periods of deformation throughout the evolution of the Appalachian Mountains. For this study, the Alleghany orogeny is of primary interest. This final mountain building event during the Permian created most of the relief seen in the central Appalachians today (Faill, 1985). The Valley and Ridge province contains the best record of Alleghanian deformation because rocks there are largely Devonian and Silurian in age. Deformation of the Valley and Ridge province can be attributed soley to the Alleghanian orogeny because these rocks are too young to record older orogenic episodes (Faill, 1985). Older rocks in the Piedmont province record a much more complex and protracted deformation history as these Ordovician and older strata record both the Alleghanian and Taconic orogenies (Faill, 1985).

PREVIOUS RESEARCH

Multiple strategies have been used to study the timing of orogenic curvature in the Pennsylvania salient. The paleomagnetic record is a reliable marker of vertical axis rotations and can be used to distinguish between primary and secondary arcs by comparing the paleomagnetic declination of equal-age rocks around the curve (Yonkee and Weil, 2009; Schwartz and Van der Woo, 1983). In a primary arc, the present day orientations of paleomagnetic declinations are parallel everywhere, while for a secondary arc paleomagnetic declinations vary systematically around the salient. Paleomagnetic data for the Pennsylvania salient reveals 20° to 30° of vertical axis rotations prior to the onset of Alleghanian folding, indicating that the salient does not fit the description of a secondary arc (Stamatakos and Hirt, 1994; Stamatakos et al., 1996).

Analysis of the structural sequence of progressive deformation is another useful tool for understanding orogenic curvature. The deformation sequence of the Valley and Ridge province has been thoroughly described. In a seminal study at the abandoned Bear Valley strip mine in Shamokin, Pennsylvania, Nickelsen (1979) recognized a six stage sequence of deformation within the Valley and Ridge province. Stage I is pre-Alleghanian jointing found only in the mine's coal seams, but stages II-VI are interpreted to represent deformation associated with the Alleghany orogeny (Nickelsen, 1979).

Others (e.g., Gray and Mitra, 1993; Spiker and Gray, 1997; Gray and Stamatakos, 1997) have since recognized the same progressive sequence of Alleghanian deformation elsewhere in the Valley and Ridge. The sequence described by Spiker and Gray (1997) is used as a reference for this study (Table 1). Overall, the sequence consists of compaction strain (labeled stage 0 because it pre-dates Alleghanian deformation) and initial layer parallel shortening (LPS) associated with top to the foreland shear (i.e., Gray and Mitra, 1993), followed by large scale flexural slip folding and finally by subsequent fold modification and late-breaking thrust faults.

The maximum shortening directions (MSDs) during subsequent phases of

Table 1: Progressive sequence of Alleghanian deformation in the Valley and Ridge Province (Spiker and Gray,1997).

Stage #	Features
0	Compaction strain
1	Grain-scale Layer Parallel Shortening: cleavage and joints
2	Layer Parallel Shortening: wedge faults and associated slickenlines
3	Flexural Slip folding: 3rd and 4th order folds, flexural slip faults and associated slickenlines
4	Late-breaking thrust faults

deformation can be determined from the spatial orientations of cleavage, joints, veins, folds and faults. Previous structural investigations from the Appalachian Plateau and Valley and Ridge show that MSDs diverge across the salient with time (Figure 1; Geiser and Engelder, 1983). In the Reading Prong, the MSDs for each stage of deformation are oriented progressively clockwise (CW) with a total of up to 30° CW rotation between the earliest and latest structures (Nickelsen, 1979; Geiser and Engelder, 1983; Gray and Mitra, 1993). Similar CW rotation has been documented farther north on the Appalachian Plateau in New York State (Geiser and Engelder, 1983; Zhou and Jacobi, 1997; Younes and Engelder, 1999). In contrast, along the Blue Ridge segment of the Pennsylvania Salient, Nickelsen (1988; 2009) documents 15-45° of counter-clockwise (CCW) rotation of shortening directions, and further to the south of the salient in western Maryland Evans (1994) and Markley and Wojtal (1996) also recognize CCW rotation. Through the hinge of the Pennsylvania salient, MSDs may remain constant in an intermediate zone of no rotation. Spiker and Gray (1997) have demonstrated this to be the case at north end of the hinge at the Appalachian front near Williamsport, PA.

Two competing kinematic models based on the observed structural data have been suggested for the formation of the Pennsylvania salient. Gray and Stamatakos (1997) have proposed a single-stage model (Figure 3A) based on data from the Appalachian Plateau and Valley and Ridge provinces. This model attributes the curvature to the geometry of a predeformational sedimentary basin. The width of a thrust belt is proportional to the thickness of sediment so that the thickest portions of the basin produce the widest portions of the belt. A salient will form centered where the basin is originally the deepest (Marshak and Wilkerson, 1992; Marshak, 2004). Additionally, the thicker center of the basin or

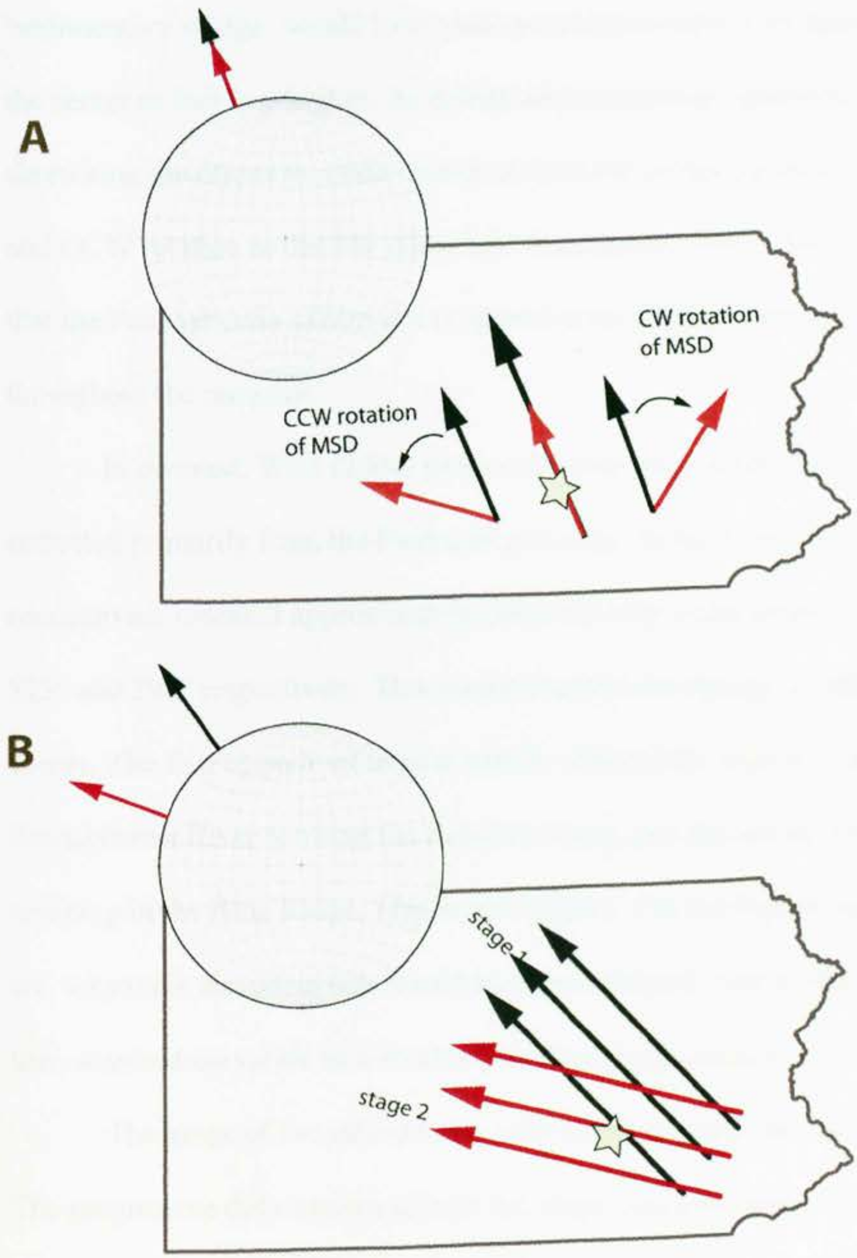


Figure 3: Schematic diagrams of the (A) single-stage and (B) two-stage models for Alleghanian tectonic shortening. Black arrows= early stage MSDs, red arrows= late stage MSDs. Stereonets show predicted outcomes at a study site (star) in the hinge of the Pennsylvania salient.

'sedimentary wedge' would have undergone more initial LPS than the thinner edges causing the center to build up higher. As deformation continued, gravitational collapse causes shortening directions to rotate outwards from the center resulting in CW rotation to the NE and CCW rotation to the SW (Gray and Stamatakos, 1997). This single-stage model predicts that the Pennsylvania salient is a progressive arc because curvature is acquired continuously throughout the orogeny.

In contrast, Wise (2004) proposed a two-stage model (Figure 3B) based on field data collected primarily from the Piedmont province. In the Piedmont, two distinct directions of transport are oriented approximately perpendicular to the limbs of the salient at azimuths 325° and 292° respectively. This model ascribes the change in MSD to two separate tectonic events. The first episode of motion mainly affected the region to the northeast of the Susquehanna River forming the Reading Prong, and the second affected the entire salient resulting in the Blue Ridge. This model implies that the Pennsylvania salient is a secondary arc, where the mountain belt would have been initially linear after Reading Prong motion but then acquired curvature as a result of the Blue Ridge motion.

The hinge of the salient is the only location where the two models obviously differ. The progressive deformation of both the single-and two-stage models would appear similar in the Blue Ridge where both call for CCW rotation of shortening directions. In the Reading prong, the single-stage model directly results in CW rotation, while the two-stage model attributes this CW rotation to re-activation of an older Taconic transport direction. However, the deformation sequence of the two models should appear very distinct in the hinge of the salient. The single stage model (Gray and Stamatakos, 1997) predicts that the hinge coincides with an axis of no rotational shortening direction and a constant MSD of ~340° (Figure 3A).

In contrast, the two-stage model (Wise, 2004) predicts two discrete shortening directions; an early stage (oriented ~325°) should be overprinted by a later stage shortening (oriented ~290°) (Figure 3B; Wise, 2004; Wise and Werner, 2004).

FIELD WORK

The so-called Clemson Island site is situated near the center of the narrow hinge of the Pennsylvania salient in the southern Valley and Ridge province of Perry County, PA (Figure 1). The Clemson Island site is a ~ 700 m long near-vertical road cut along PA Route 11/15 that exposes the Irish Valley member of the Devonian Catskill Formation. Alternating layers of coarser grained sandstone and finer grained siltstone and shale reveal a sequence of Alleghanian deformation very similar to that documented by Spiker and Gray (1997) ~100 km northwest near Williamsport, PA.

A detailed structural investigation of the outcrop was completed over a three week field season. The upper parts of the outcrop were accessible in some spots, but in general, structural measurements could only be obtained in the lowest 2-3 m after brush had been cleared away using a machete, hedge clippers and a chainsaw. Structural measurements of the orientations of planar and linear features associated with each stage of deformation were noted. Stage 1 grain-scale LPS is manifested in the outcrop by planar cleavage and joints. Wedge faults and associated slickenlines are the dominant manifestation of stage 2. Measureable features of stage 3 folding include bedding orientations, flexural slip faults and their associated slickenlines, and a pencil cleavage. The final stage 4 is characterized by late breaking thrust faults. Each structural measurement was paired with the present day bedding orientation. This enables features to be restored to their orientations at the time of formation.

Two oriented samples showing cleavage-bedding intersection lineation were also collected in the field for grain scale analysis: sample A from the northern end of the outcrop and sample B from the southern end of the outcrop (Figure 4).

LAB WORK

Structural features form in specific orientations with respect to the MSD and so stereonets can be used to quickly constrain the orientations of MSDs associated with the formation of each type of feature. LPS cleavage, wedge faults, thrust faults and folded bedding all form striking perpendicular to the MSD. Joints often occur in orthogonal sets with one joint plane oriented perpendicular to the MSD and orthogonal joint plane parallel to the MSD.

The orientation measurements of all structural features were plotted on lower hemisphere equal area stereonets. Linear features were plotted as lines and planar features as poles to the planes. For the first two stages of deformation (those occurring prior to folding), bedding was restored to horizontal in order to rotate all these features back to pre-folding orientations. Pre-folding features cluster more tightly on stereonets when bedding is restored to horizontal, whereas post-folding features cluster more tightly plotted in their present day orientations. This "fold test" serves as a double check for placing features correctly into the sequence of deformation.

Grain-scale deformation can be measured using the normalized Fry method (Fry, 1979; Erslev, 1988). This method is based on the premise that the distances between the centers of adjacent grains in a deformed rock can be used to quantify the strain that acted on that rock, so long as the centers were originally evenly distributed (Ramsay, 1967). The Fry method involves situating a single point of an overlay over the centers of many grains and

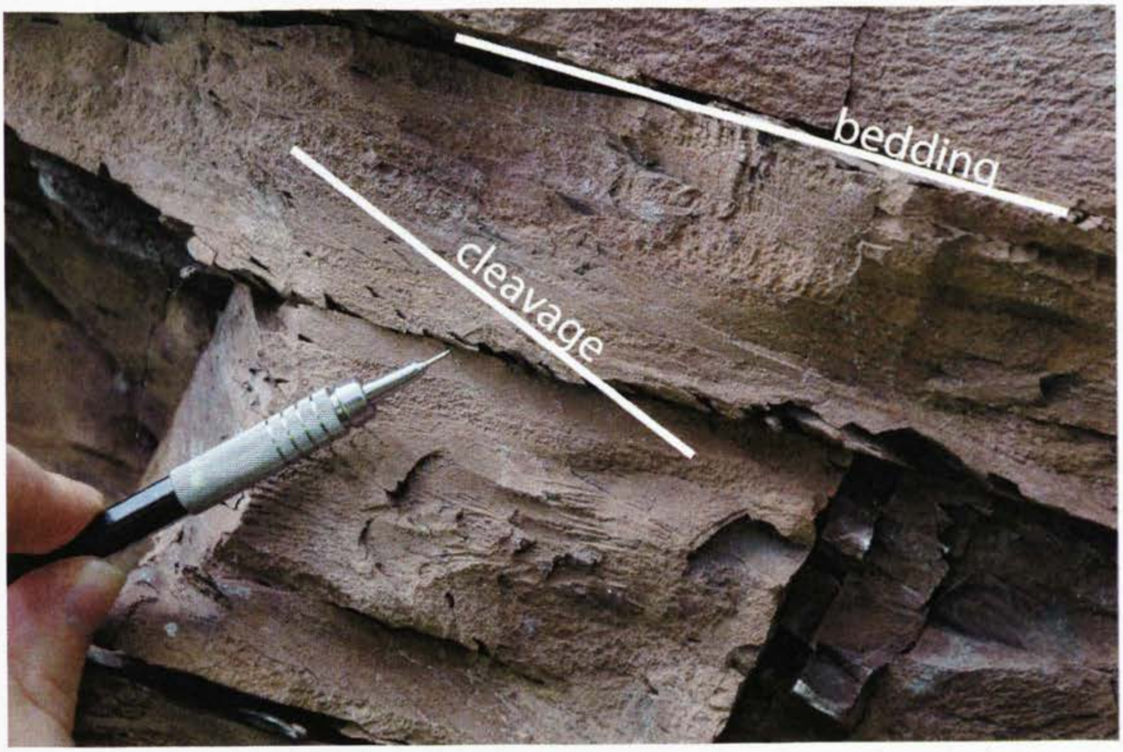


Figure 4: Oriented Sample A was taken from the bed just above the pencil in this photo. The pencil points to the cleavage bedding intersection lineation.

recording the positions of adjacent centers as dots on the overlay. The result is a dense dot field surrounding an elliptical void space which represents the finite strain ellipse for the deformed rock (Fry 1979). The normalized Fry method accounts for original heterogeneity in the undeformed rock by dividing the distance between any two centers by the sum of the grains' radii. This removes inaccuracies due to variations in grain size and poor sorting, resulting in a more sharply defined ellipse (Erslev, 1988).

To constrain grain-scale deformation at Clemson Island, three mutually perpendicular thin sections were cut from sample A: one oriented parallel to the measured cleavage-bedding intersection lineation, another parallel to this lineation and a third mutually perpendicular to the other two. Six oriented thin sections were cut from sample B. Three were cut respect to the cleavage-bedding intersection lineation, as described above for sample A. The remaining three were cut with respect to bedding; one oriented parallel to bedding, one perpendicular to bedding and a third mutually perpendicular to the other two. A Nikon P400 petrographic microscope equipped with an Insight Spot digital camera was used to collect photomicrographs of each thin section. Because grain boundaries may not be entirely apparent at a single orientation under cross-polarized light, a series of three photomicrographs were taken for each thin section at 30° intervals of rotation and these were aligned and overlain to create a composite image. These composite images have clearly defined grain boundaries and were used as the photographic bases for constraining the grain scale deformation.

The normalized Fry method was carried out by using Matlab. The centers, long axes, and short axes of about 100-150 grains were measured for each image, and the computer program generated finite strain ellipses from this information (Figure 5). Measurements of

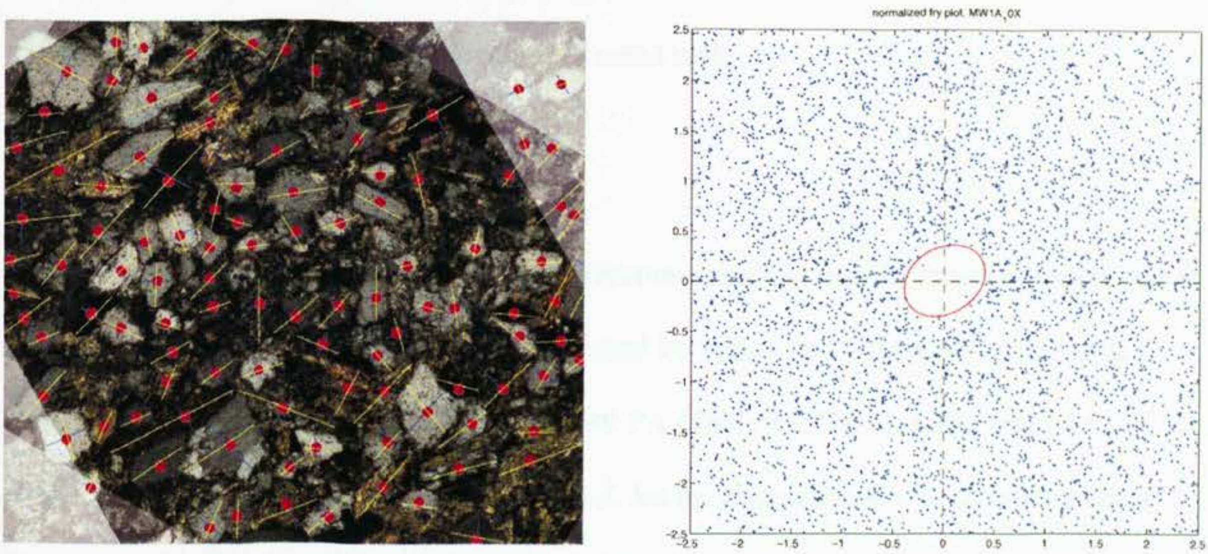


Figure 5: An oriented photomicrograph (sample MW1A) used to calculate the finite strain ellipse (R=1.3) by the normalized Fry method.

the finite strain ellipses on three orthogonal surfaces were then combined to obtain the three dimensional finite strain ellipsoid for the deformed rock.

RESULTS

Cross-cutting relationships between structures confirm that the structural sequence at Clemson Island mimics the sequence documented by Spiker and Gray (1997) along the Alleghany front in the vicinity of Williamsport, PA (Table 1). The cleavage and joints of stage 1 are cross-cut by wedge faults of stage 2. Joints, cleavage and wedge faults cluster better when bedding is restored to horizontal, indicating that these stage 1 and 2 structures have been subsequently re-oriented by stage 3 folding. Multiple joint surfaces containing slickenlines and some wedge faults with two non-parallel sets of slickenlines are indicators that these surfaces have been subsequently re-activated as fault surfaces and pre-date stage 3 folding. Finally, the thrust faults of stage 4 cross-cut all other features in the outcrop.

Stage 1: Grain scale LPS

Finite Strain

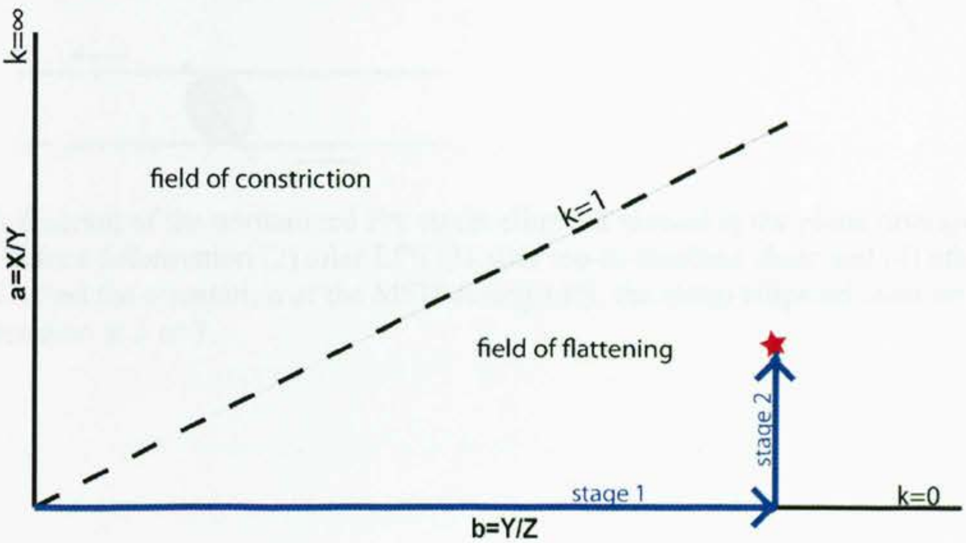
Finite strain measures the total deformation in a rock between its initial and final states but is independent of the incremental strain path in between (Van der Pluijm and Marshak, 2004). Grain-scale finite strain was quantified by the normalized Fry method for three mutually perpendicular thin sections oriented with respect to bedding. The axes of the finite strain ellipsoid at Clemson Island are X= 1.14, Y= 1.04, Z= 0.84 (where X, Y, Z correspond to the major, intermediate, and minor axes respectively), assuming no volume loss during deformation.

Finite strain can be represented on a 2-dimensional Flinn Diagram plot (Figure 6). The ratio of the maximum stretch to the intermediate stretch (a=X/Y) is plotted on the y-axis and the ratio of the intermediate stretch to the minimum stretch (b= Y/Z) on the x-axis. The plot is divided into two fields by the line k=1 representing plane strain, where k= (a-1)/(b-1). Strain ellipsoids that plot on the a-axis (k=∞) represent pure uniaxial extension and ellipsoids that plot on the b axis (k=0) represent pure flattening. The field of flattening is where k<1, and the field of constriction where k>1 (Van der Pluijm and Marshak, 2004). For the Clemson Island site, the calculated values of a=1.10, b=1.24, and k=0.42 of the strain ellipsoid fall in the field of flattening strain.

The Flinn diagram can also be used infer the strain path that the rock underwent to achieve the finite strain ellipsoid. The vector representing finite strain can be broken into XY plane and YZ plane components. Spiker and Gray (1997) observe flattening of crinoid ossicles as evidence for compaction strain in the XY plane that predates strain in the YZ plane. YZ plane deformation is attributed to grain-scale LPS. Although there are no crinoid ossicles as markers of compaction strain at Clemson Island, the same strain path of compaction followed by LPS is inferred. The MSD of grain-scale LPS can also be found from the orientation of the major axis of the strain ellipse parallel to the bedding plane when bedding and the ellipse are restored to their original orientations before folding. In this case, the MSD is 335° (Figure 7).

Joints

Joints were recognized by Nickelsen (1979) as the first Alleghanian features to appear in the deformation sequence at Bear Valley. For this reason, they have been placed into stage 1 of the deformation sequence at Clemson Island. Joints are prevalent in the sandstone layers



Fig

Figure 6: Flinn diagram where a is the ratio of the long axis to the intermediate axis, and b is the ratio of the intermediate axis to the short axis of the finite strain ellipsoid. K=(a-1)/(b-1). The red star represents the finite strain ellipsoid from normalized Fry analysis of the study site. The blue arrows show the strain path to the finite strain ellipsoid.

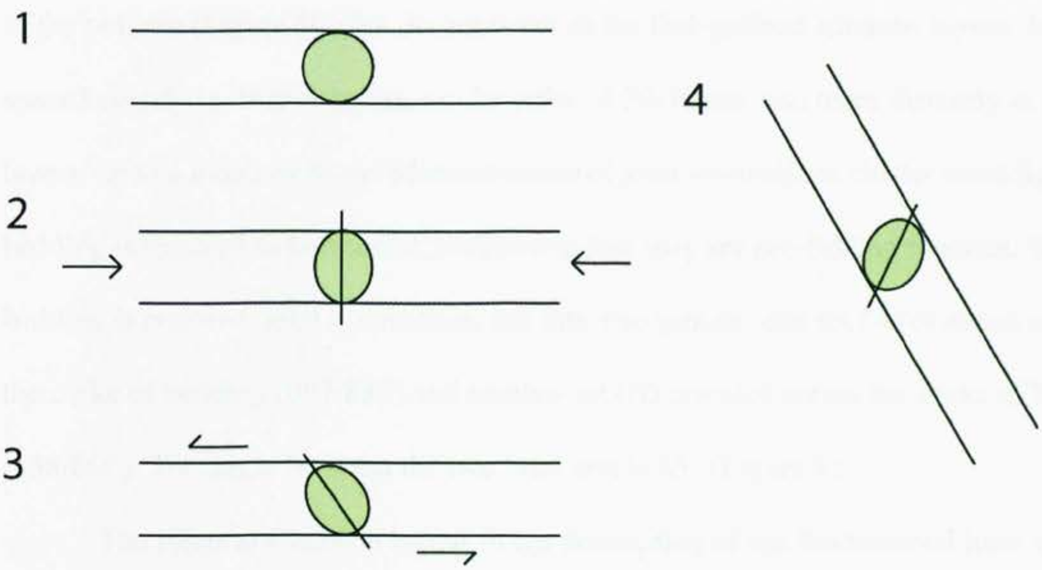


Figure 7: diagram of the normalized Fry strain ellipsoid viewed in the plane orthogonal to S_1/S_0 (1) before deformation (2) after LPS (3) after top-to-foreland shear and (4) after folding. To find the orientation of the MSD during LPS, the strain ellipsoid must be restored to its orientation at 2 or 3.

of the outcrop (Figure 8C) but do not occur in the fine-grained siltstone layers. Joints are spaced closely in thinner layers, on the order of 20-30 cm, and more distantly in thicker layers, up to a meter or more. Measurements of joint orientations cluster more tightly when bedding is restored to horizontal, confirming that they are pre-folding features. When bedding is restored, joint orientations fall into two groups: one set (A) oriented subparallel to the strike of bedding (067/83S) and another set (B) oriented across the strike of bedding (158/86E). The angle between the two joint sets is 85° (Figure 9).

The joints at Clemson Island fit the description of the fundamental joint system identified by Nickelsen and Hough (1967) in the coals and shales on the Appalachian Plateau. The system consists of two orthogonal sets of joints; one systematic set and one non-systematic set. Systematic joints are laterally extensive, planar surfaces interpreted as extension fractures that form perpendicular to the direction of least compressive stress under the presence of fluid pressure. They are typically oriented perpendicular to bed boundaries and across the strike of bedding. Non-systematic joints are more irregular curved surfaces that span the intervals between systematic joints and are truncated by them. They are oriented along the strike of bedding (Hough, 1961) and are believed to be "release fractures" that form perpendicular to the greatest principal stress once it is released (Nickelsen and Hough, 1967).

Set A joints are along-strike joints, which can be interpreted as extension fractures oriented perpendicular to the direction of minimum shortening. Therefore their MSD during formation is the azimuth of the great circle defining the plane of best fit to the poles to these joints when they are plotted with S_0 restored to horizontal. This azimuth is ~340°. Set B joints are the cross-strike set oriented perpendicular to the MSD. Therefore, the MSD during

Figure 8: Photos of structural features in the outcrop. **A:** S_1 cleavage. **B:** a joint surface with slickenlines indicating it has been re-activated as a fault surface **C:** orthogonal joint planes **D/E:** conjugate wedge faults with arrows indicating slip directions.

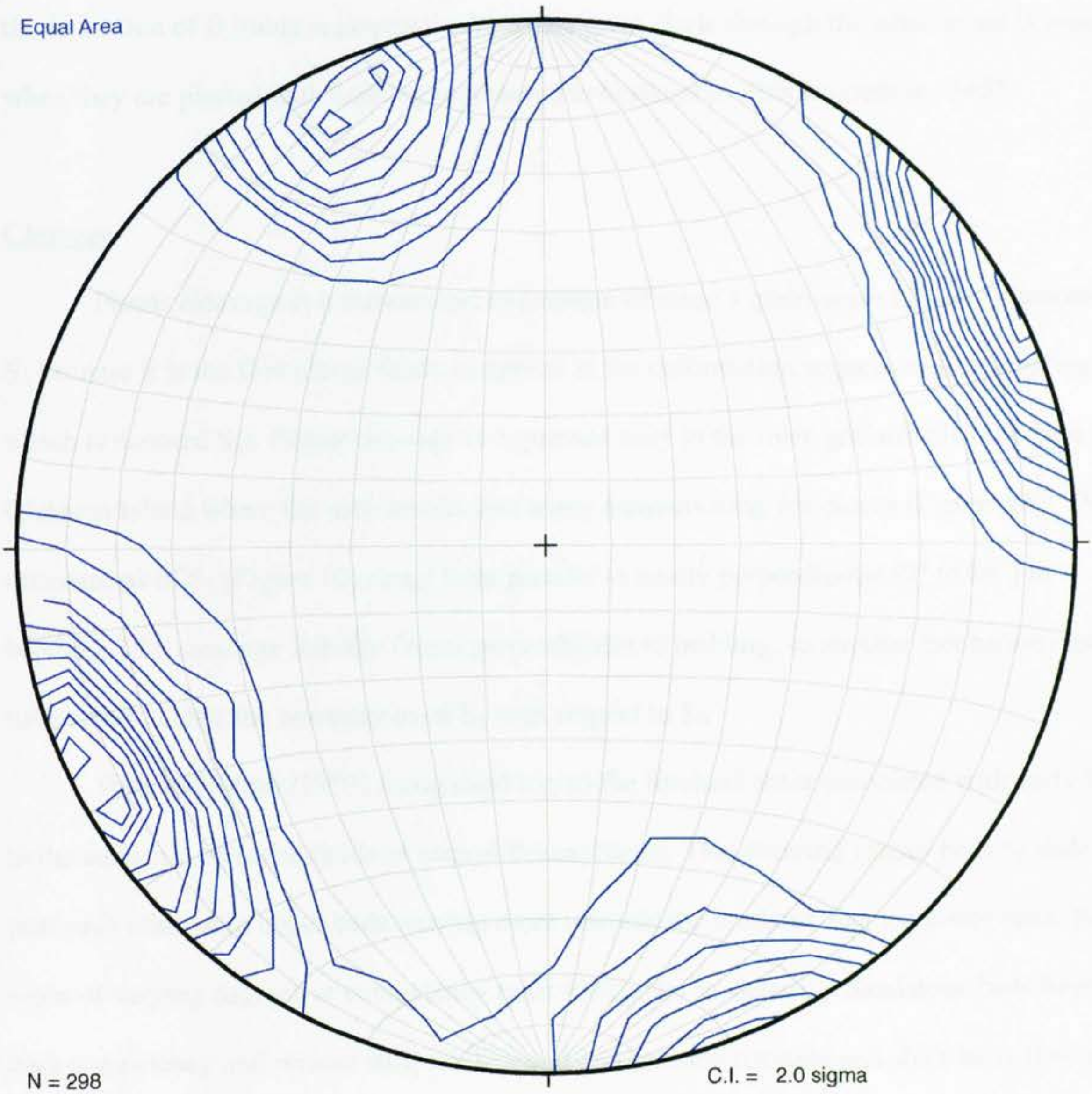


Figure 9: Kamb plot of poles all joint plane measurements, plotted restored to their original orientations before folding. The joints define two sets with average orientations of 067/83S and 158/86E.

the formation of B joints is perpendicular to the great circle through the poles to set B joints when they are plotted with bedding unrestored to horizontal. This azimuth is ~345°.

Cleavage

Planar cleavage is a mesoscopic expression of stage 1 grain-scale LPS. It is denoted S_1 because it is the first planar fabric to appear in the deformation sequence (after bedding which is denoted S_0). Planar cleavage is expressed only in the finer-grained siltstone beds at Clemson Island where the rock breaks into many anastamosing flat pieces (Figure 8A). The orientations of S_1 (Figure 10) range from parallel to nearly perpendicular (0° to 88°) to bedding. LPS cleavage initially forms perpendicular to bedding, so another mechanism must have acted to alter the orientations of S_1 with respect to S_0.

Gray and Mitra (1993) recognized top-to-the foreland shear associated with early LPS in the eastern Anthracite fields of central Pennsylvania. This shearing causes beds to slide past each other with upper beds moving more towards the foreland than the lower ones. Rock types of varying degrees of competency react differently to shearing. Sandstone beds have high competency and remain stiff, while lower competency siltstone and shale beds flow and acquire shear strain (Van der Pluijm and Marshak, 2004). During top-to-the-foreland shear, the cleavage in siltstone and shale beds at Clemson Island was likely reoriented as sandstone beds above and below moved rigidly. Cleavage in these siltstone and shale beds is often oriented at a high angle to bedding in the center of beds but transitions to a lower angle at the edges where sandstone layers slid past.

Top-to-the-foreland shearing can also be observed by measuring the strain ellipse on the plane orthogonal to the S_1/S_0 intersection lineation (Figure 11). For both samples, the

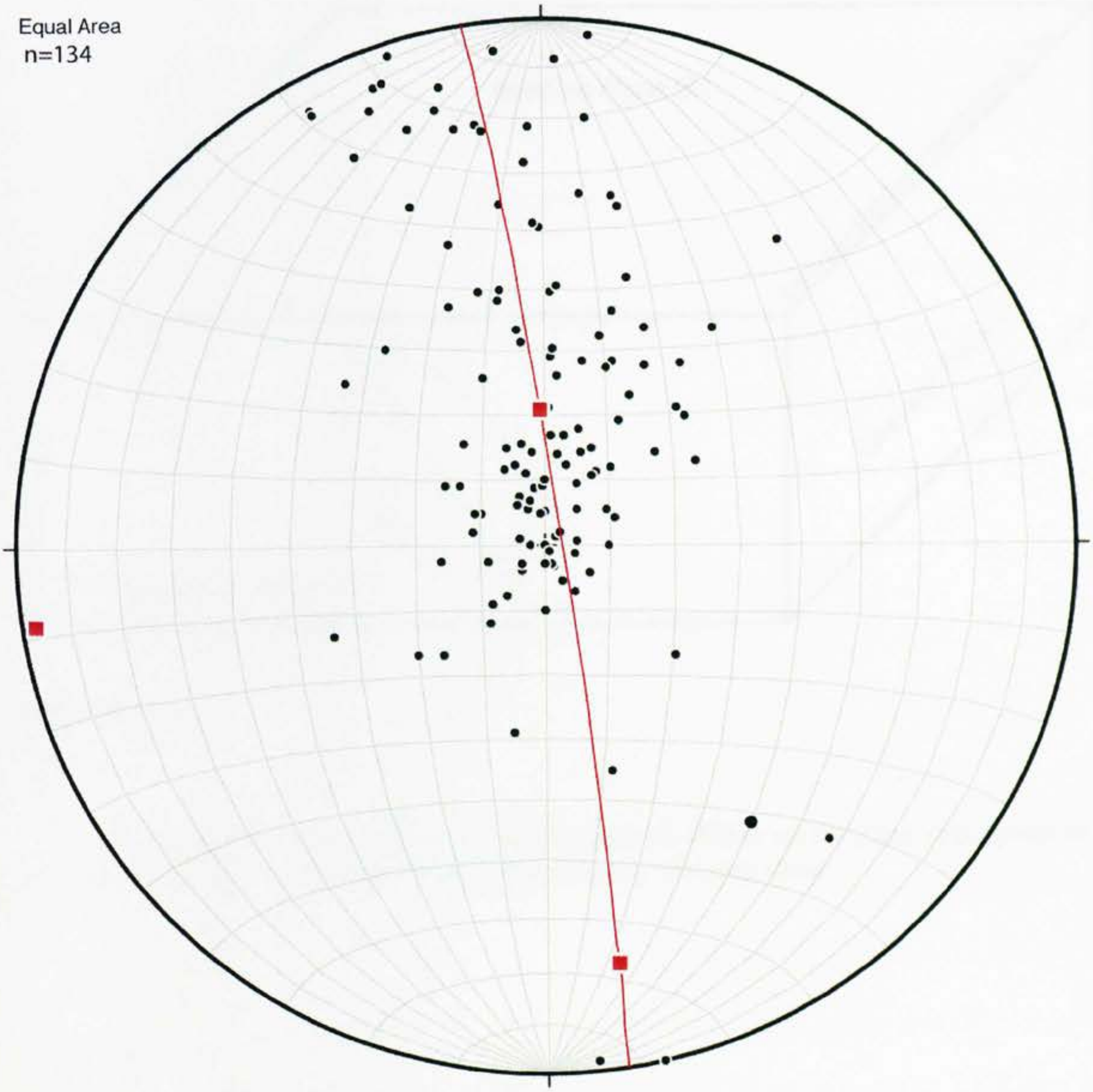


Figure 10: Poles to all S1 measurements, plotted in their orientations at time of formation, with bedding restored to horizontal. Red line is the line of best fit to the poles, and also the aziumuth of the MSD during the formation of S1, ~351°

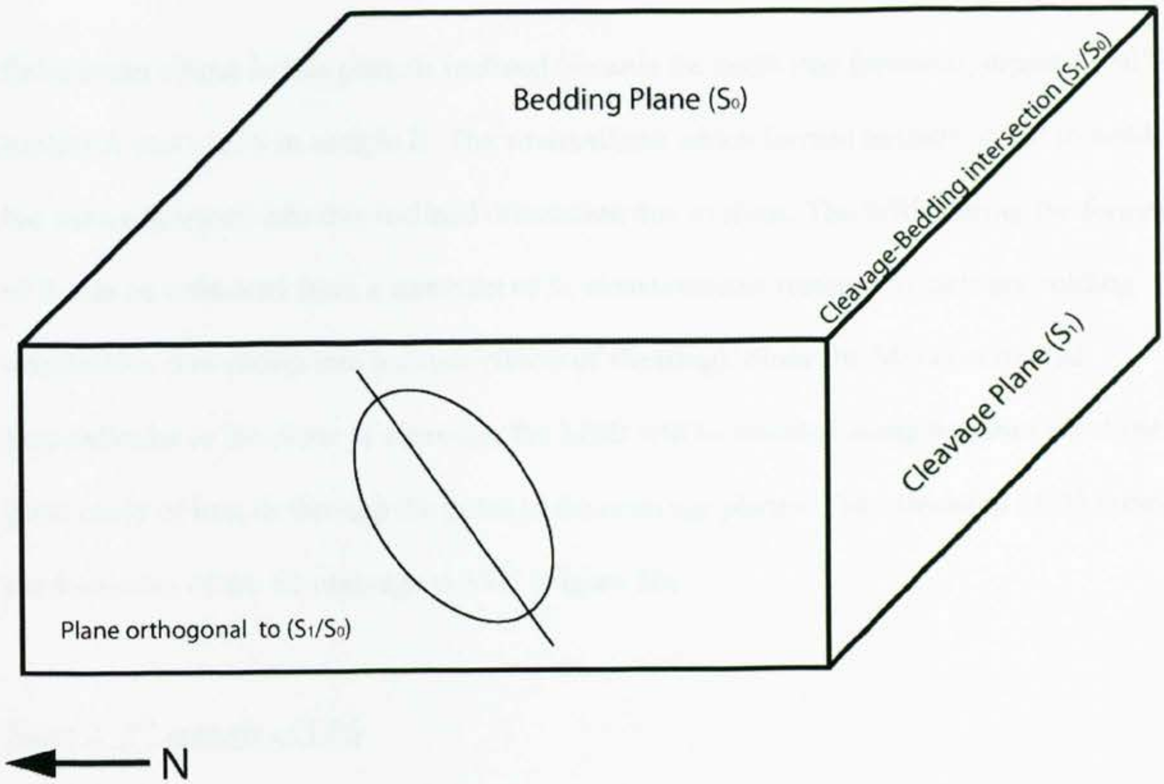


Figure 11: Diagram showing an inclined strain ellipse on the plane orthogonal to the cleavage-bedding (S_1/S_0) intersection, indicating top-to-north shear.

finite strain ellipse in this plane is inclined towards the north (the foreland), dipping ~50°S in sample A and ~55°S in sample B. The strain ellipse which formed initially at 90° to bedding has been reoriented into this inclined orientation due to shear. The MSD during the formation of S_1 can be estimated from a stereonet of S_1 measurements restored to their pre-folding orientations (not taking into account effects of shearing). Since the MSD is oriented perpendicular to the plane of cleavage, the MSD will be oriented along the azimuth of the great circle of best fit through the poles to the cleavage planes. The calculated MSD during the formation of the S1 cleavage is 351° (Figure 10).

Stage 2: 2^{nd} episode of LPS

Wedge Faults

LPS is accommodated in competent beds by the development of wedge faults which result in thickening of the faulted bed as it is shortened. Wedge faults are common throughout the Valley and Ridge province (Mitra, 2002) and are the main expression at the Clemson Island outcrop of a second episode of LPS. Wedge faults are commonly recognized within individual beds and across multiple beds. Often wedge faults occur in conjugate sets (Figure 8 D/E). The fault planes are ornamented with slickenlines indicative of dip parallel slip. In some instances the dip-parallel slickenlines are overprinted by a second set of slickenlines indicative of fault reactivation during later stages of deformation.

Stage 2 wedge faults have been passively rotated during stage 3 folding. Consequently, to determine the orientation of the MSD when the wedge faults formed requires unfolding the folds on a stereonet. Wedge faults strike perpendicular to the MSD and their down-dip slickenlines trend parallel to the direction of shortening (Mitra, 2002).

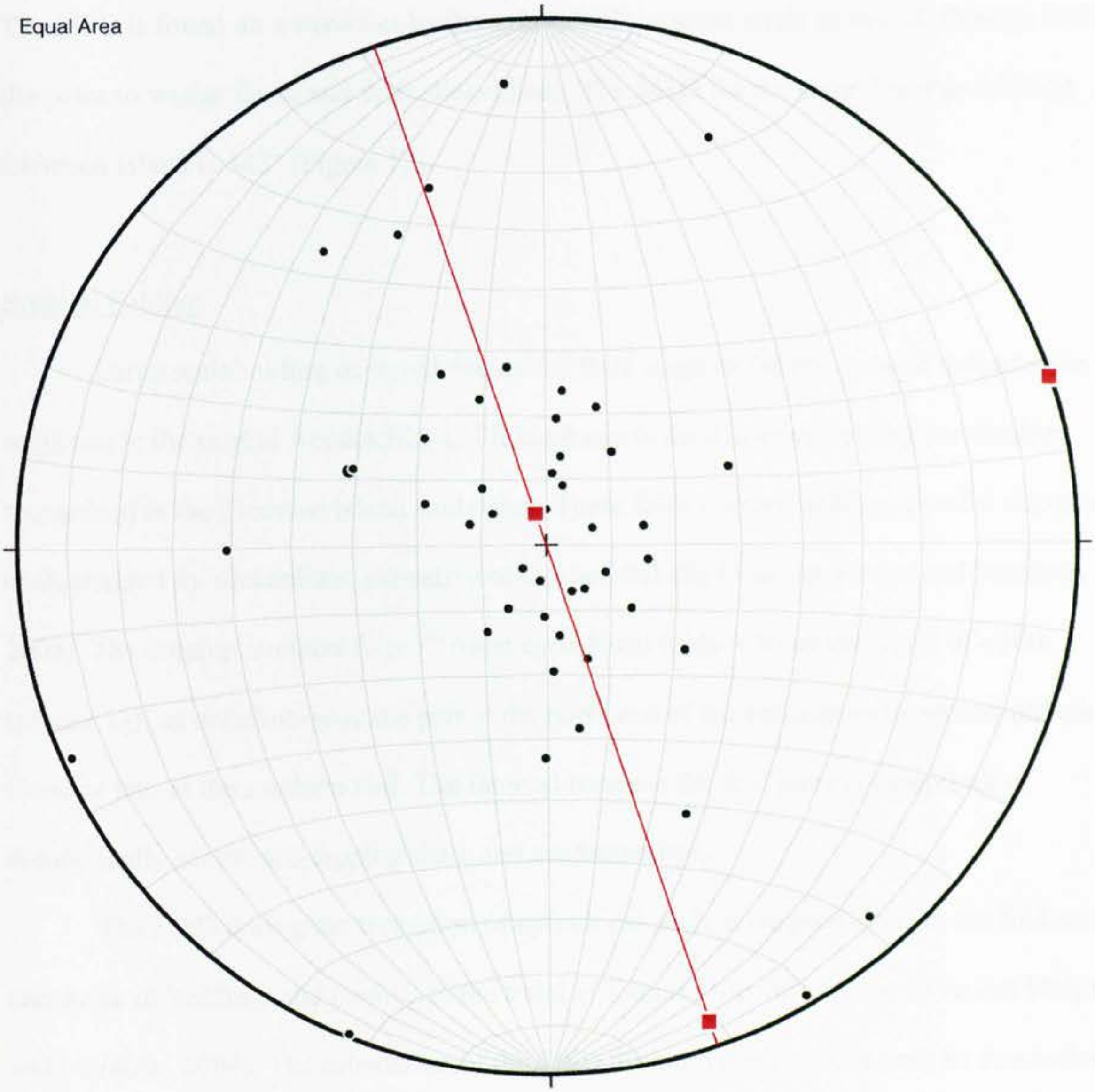


Figure 12: Poles to wedge faults, with S_0 restored to horizontal.

The MSD is found on a stereonet by the azimuth of the great circle of best fit through both the poles to wedge faults and their slickenlines. The MSD for the stage 2 wedge faults at Clemson Island is 342° (Figure 12).

Stage 3: Folding

Large scale folding occurred during the third stage of the progressive deformation sequence in the central Appalachians. Flexural slip is the dominant folding mechanism recognized at the Clemson Island study area. These folds contain bedding-parallel slip planes characterized by slickenlines indicative of dip-parallel slip (Van der Pluijm and Marshak, 2004). The outcrop contains four 3rd order cylindrical folds with wavelengths of ~10m (Figure 13); an anticline-syncline pair at the north end of the outcrop and a second anticline-syncline pair at the southern end. The interval between the fold pairs is comprised of homoclinally southward-dipping shale and sandstone beds.

The MSD during the formation of cylindrical folds is perpendicular to the fold axes and strike of bedding, and parallel to the trend of flexural slip slickenlines (Van der Pluijm and Marshak, 2004). The orientation of the MSD during flexural folding can be constrained via two independent methods. On a stereonet of the poles to bedding, the azimuth of the great circle of best fit to the poles defines the MSD, in this instance the MSD is 340° (Figure 14). Alternatively, the MSD can be determined by fitting a great circle of best fit through the poles to the flexural slip fault planes and associated slickenlines, yielding an MSD orientation of 339° (Figure 15).

Pencil cleavage is another expression of folding in the outcrop. The rock in fine grained beds in the cores of folds breaks into long, narrow "pencils" ranging from ~5-20 cm

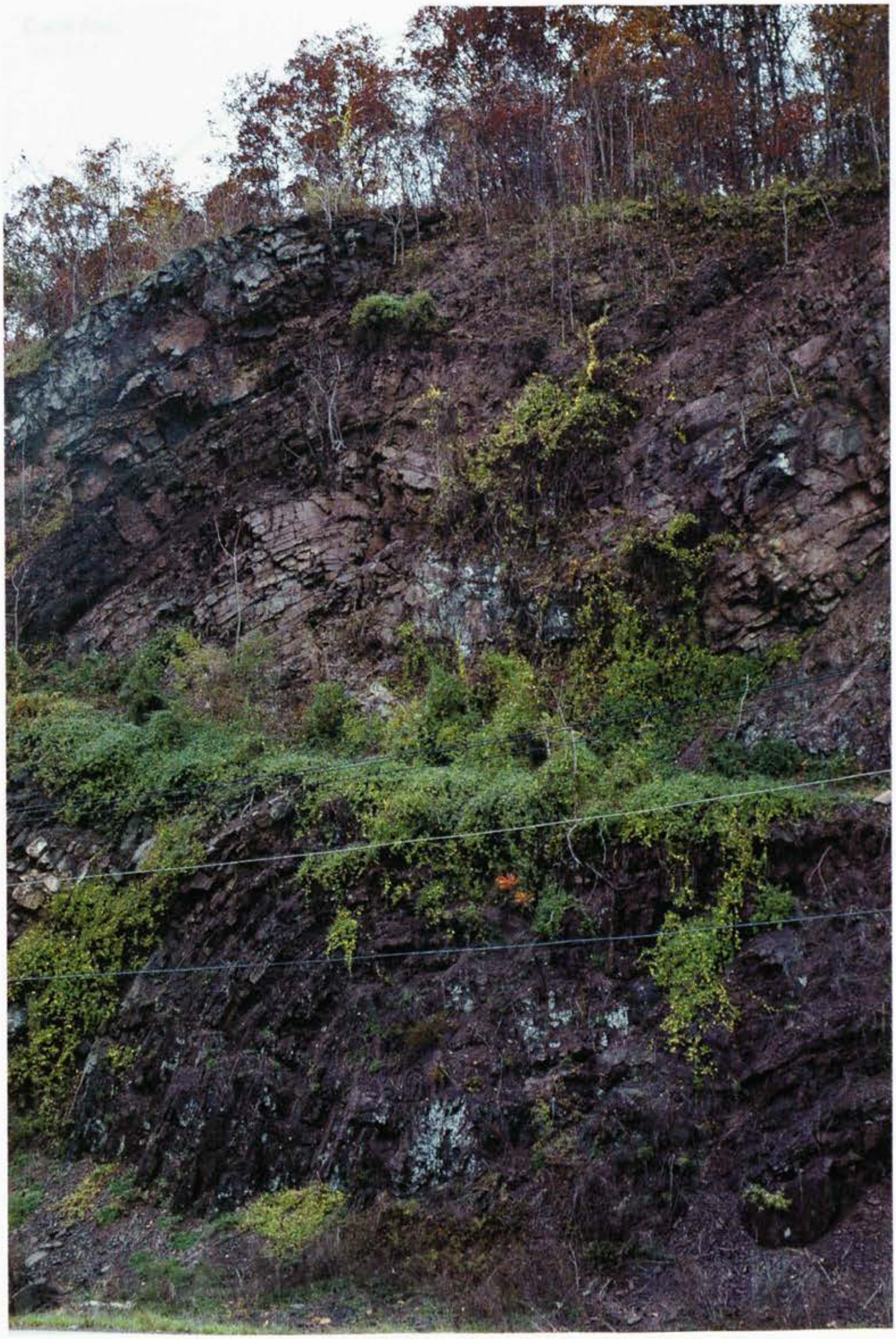


Figure 13: Photo of a 3[rd]-order anticline at the southern end of the outcrop

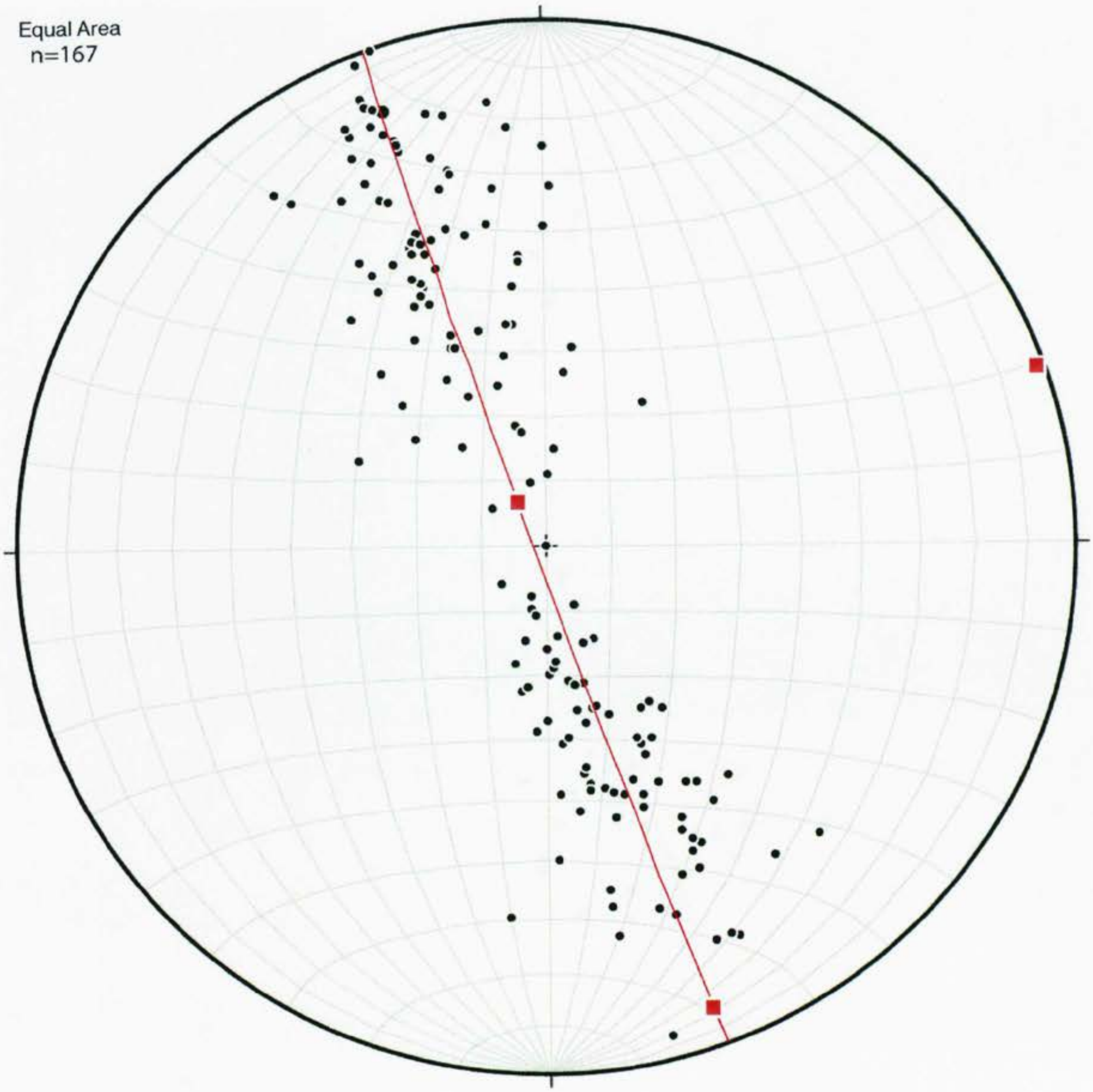


Figure 14: Poles to bedding measurements around all major outcrop folds. Includes data from this study as well as data collected during the fall of 2008 by Jake Davidson, Gwen Dunnington and Jake Davidson. Red line is the best fit to the poles and also represents the MSD during the formation of the folds, ~340°

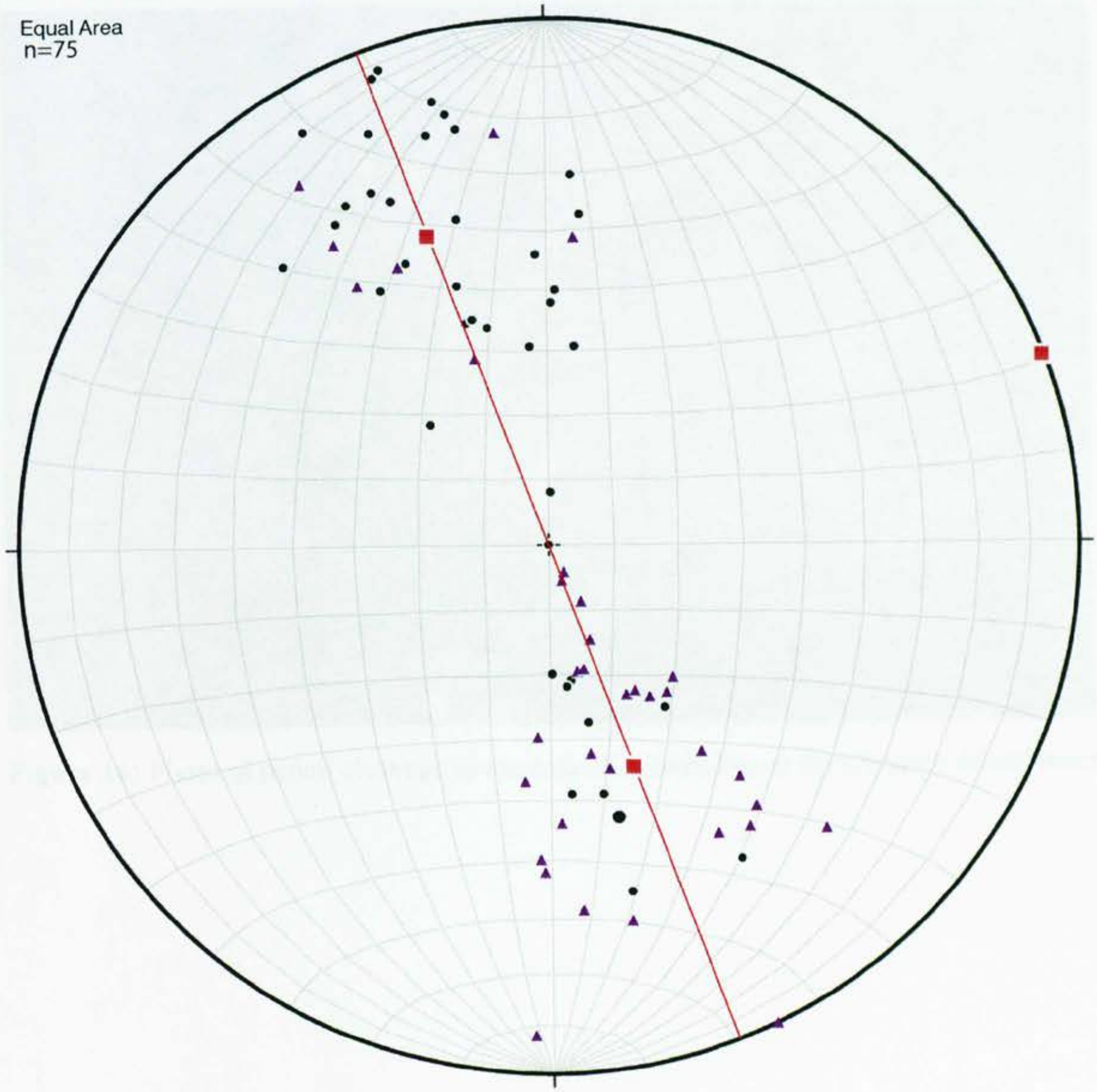


Figure 15: Poles to all flexural slip faults (black dots) and flexural slip slickenlines (purple triangles). Red line is the best fit to both, and represents the MSD during the formation of the flexural slip faults, ~340°.

Figure 16: Photo of pencil cleavage in the core of an anticline at the Clemson Island outcrop.

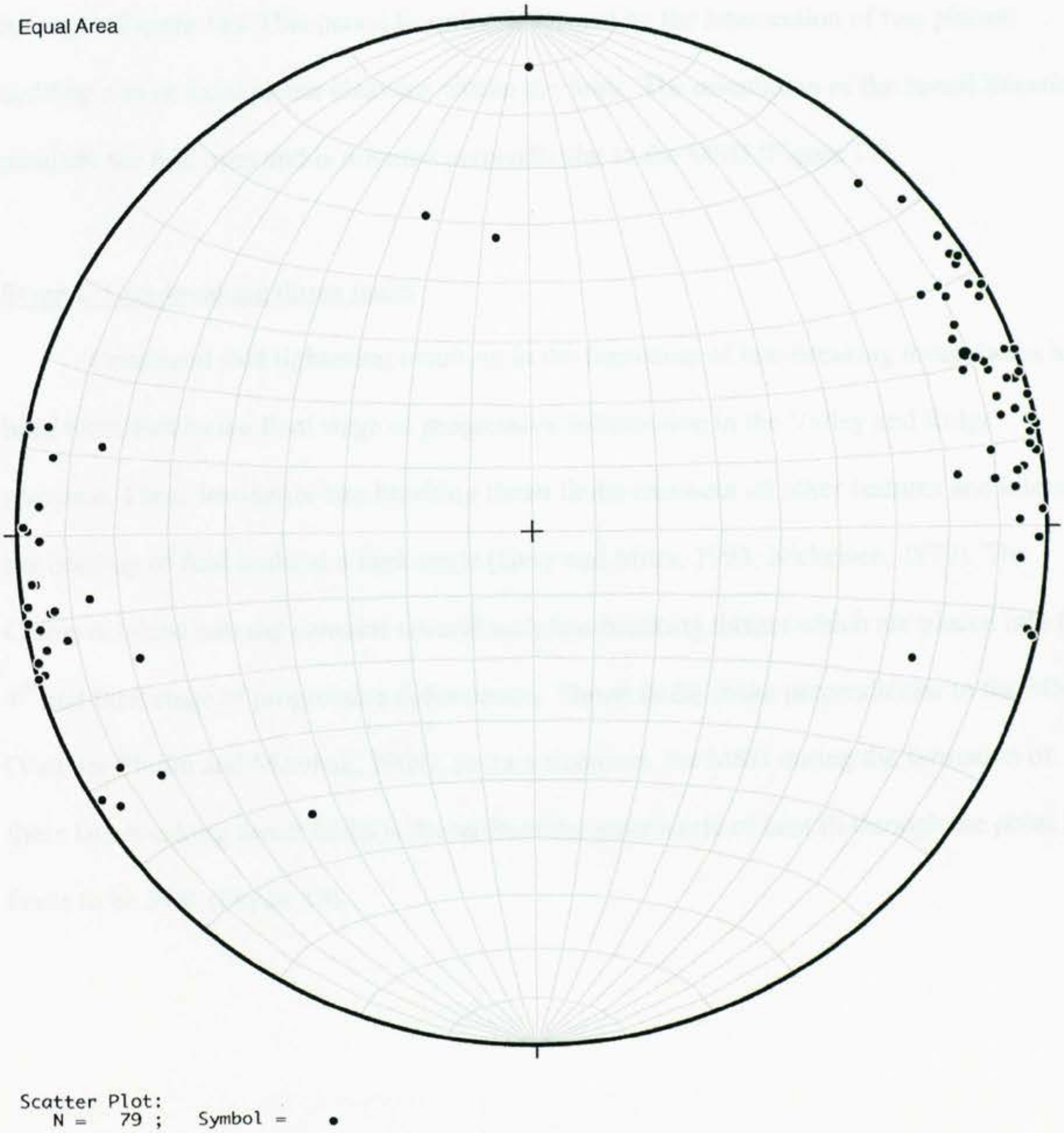


Figure 17: All pencil cleavage measurements plotted with bedding restored to horizontal.

in length (Figure 16). This pencil lineation is formed by the intersection of two planes; bedding and an axial planar cleavage within the folds. The orientation of the pencil lineation parallels the fold axes and is oriented perpendicular to the MSD (Figure 17).

Stage 4: Late-breaking thrust faults

Continued fold tightening resulting in the formation of late-breaking thrust faults has been identified as the final stage of progressive deformation in the Valley and Ridge province. These low-angle late-breaking thrust faults cross-cut all other features and intersect the bedding of fold limbs at a high angle (Gray and Mitra, 1993; Nickelsen, 1979). The Clemson Island outcrop contains several such late-breaking thrusts which are placed into the 4th and final stage of progressive deformation. Thrust faults strike perpendicular to the MSD (Van der Pluijm and Marshak, 2004), so on a stereonet, the MSD during the formation of these late-breaking thrust faults is found from the great circle of best fit through the poles to faults to be 343° (Figure 18).

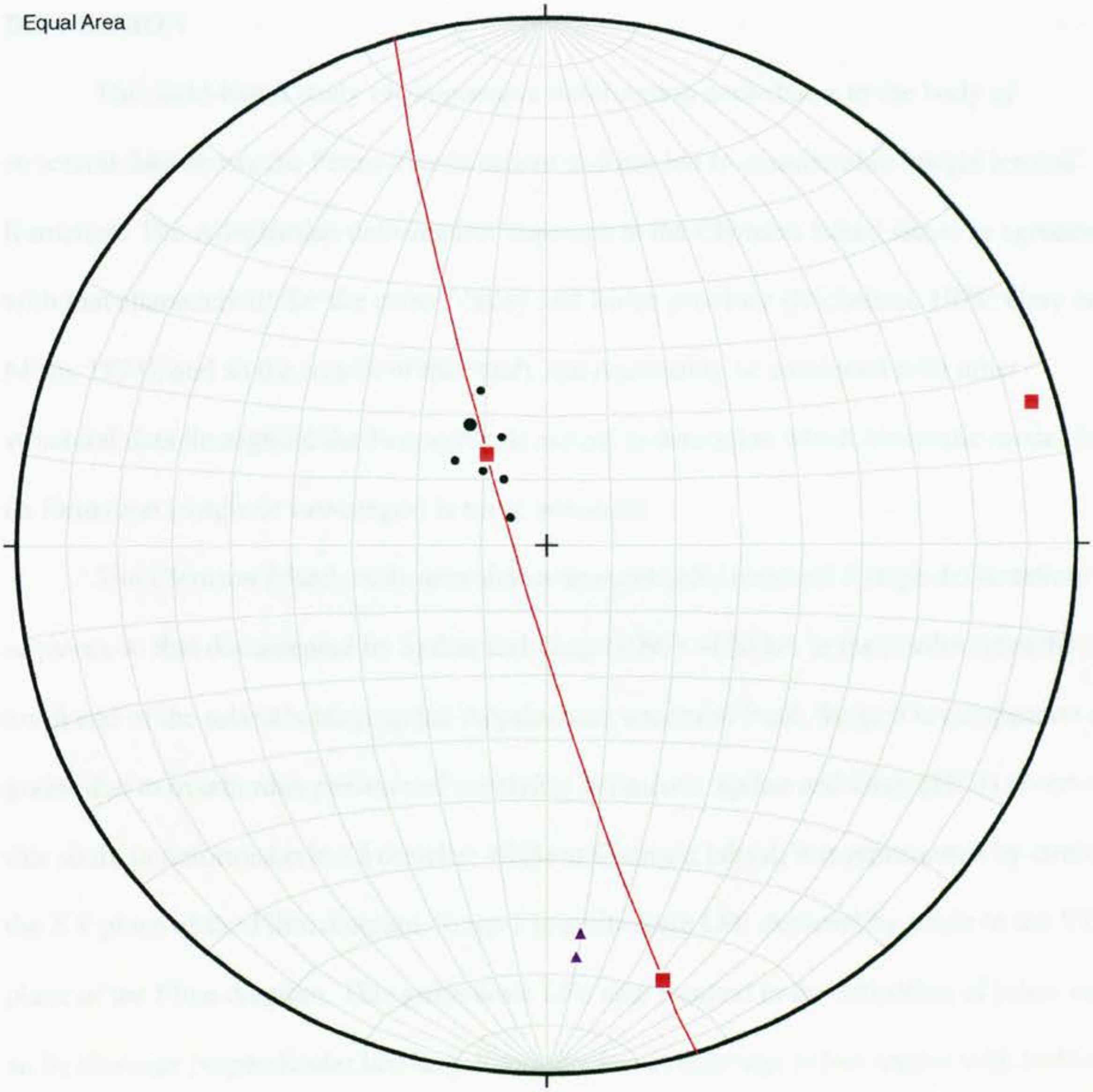


Figure 18: Poles to all late-breaking thrust faults. Red line of best fit represents the MSD during the formation of the thrust faults, ~343°.

DISCUSSION

This field-based study of progressive deformation contributes to the body of structural data across the Pennsylvania salient and has led to considerable insight into its formation. The Alleghanian deformation sequence at the Clemson Island site is in agreement with that characteristic for the entire Valley and Ridge province (Nickelsen, 1979; Gray and Mitra, 1993), and so the results of this study can reasonably be combined with other structural data throughout the Pennsylvania salient to determine which kinematic model for its formation (single or two-stages) is more accurrate.

The Clemson Island study area shows an essentially identical 5 stage deformation sequence to that documented by Spiker and Gray (1997) ~100 km to the northwest at the north end of the salient's hinge at the Appalachian structural front. Stage 0 is compaction of grains due to overburden pressure of overlying sediments. Spiker and Gray (1997) observe this strain in deformed crinoid ossicles, while at Clemson Island, it is represented by strain in the XY plane of the Flinn diagram. Stage 1 is grain-scale LPS depicted by strain in the YZ plane of the Flinn diagram. This grain-scale LPS also resulted in the formation of joints and an S_1 cleavage perpendicular bedding. Reorientation of cleavage to low angles with bedding and a strain ellipse inclined to the north on the plane orthogonal to bedding suggest that stage 1 structures were affected by top-to-the-foreland shearing. This shearing is contemporaneous with stage 1 LPS, as suggested by Gray and Mitra (1993). Stage 2 of the progressive deformation sequence at Clemson Island is a renewed episode of LPS which is manifested mesoscopically by conjugate wedge and wrench faults in the outcrop which cross-cut stage 1 structures. Stage 3 folding occurred by flexural slip mechanism and rotated the preexisting structures, and finally late-breaking thrust faults formed last as a result of continuing fold tightening during stage 4.

The MSDs recorded by all stages of Alleghanian deformation at the Clemson Island site vary in orientation by less than 20°. They range from 335° to 351° with no systematic change over time either CW or CCW (Table 2). The average MSD for all Alleghanian deformation is 342°. This constant MSD at the southern end of the salient's hinge is essentially identical to the findings of Spiker and Gray (1997) at the hinge's northern end at the Appalachian structural front, where a consistent MSD of 341° was found (Figure 19). Together, these investigations show that the hinge of the Pennsylvania salient coincides with an axis of no rotation. This observation is in agreement with the single-stage model proposed by Gray and Stamatakos (1997) which predicts a single MSD of ~340° for the entire Alleghanian deformation sequence through the hinge of the Pennsylvania salient.

These results are inconsistent with the two-stage model proposed by Wise (2004). The two-stage model is based on data from the Piedmont province where two distinct directions of shortening are observed, the first with an MSD oriented ~325° and the second oriented ~290°. This two-stage model uses the observed Piedmont data to explain deformation in the Valley and Ridge province, relating the first stage of motion to the formation of the Reading Prong and the second to the Blue Ridge. Structural data from the Blue Ridge are in agreement with the two-stage model because MSDs to the south show CCW rotation from MSDs in the hinge and parallel the Piedmont MSDs of ~290°. However, the two-stage model predicts that the hinge of the Pennsylvania salient should reflect MSDs associated with the formation of both the Reading Prong and the Blue Ridge. An early MSD oriented ~325° should be overprinted by a later MSD oriented ~290°. This is distinctly

Table 2: MSDs calculated from structural features for each stage of the deformation sequence in this study compared to those found by Spiker and Gray (1997).

Stage #	Feature	This Study	Spiker and Gray, 1997
0	Compaction strain		
1	Finite strain ellipsoid	335	346
	S1	351	339
	Joints (set A)	340	340
	Joints (set B)	345	329
2	Wedge faults	342	338
3	Folds	340	340/ 341
	Flexural slip faults	339	341
4	Late thrust faults	343	351

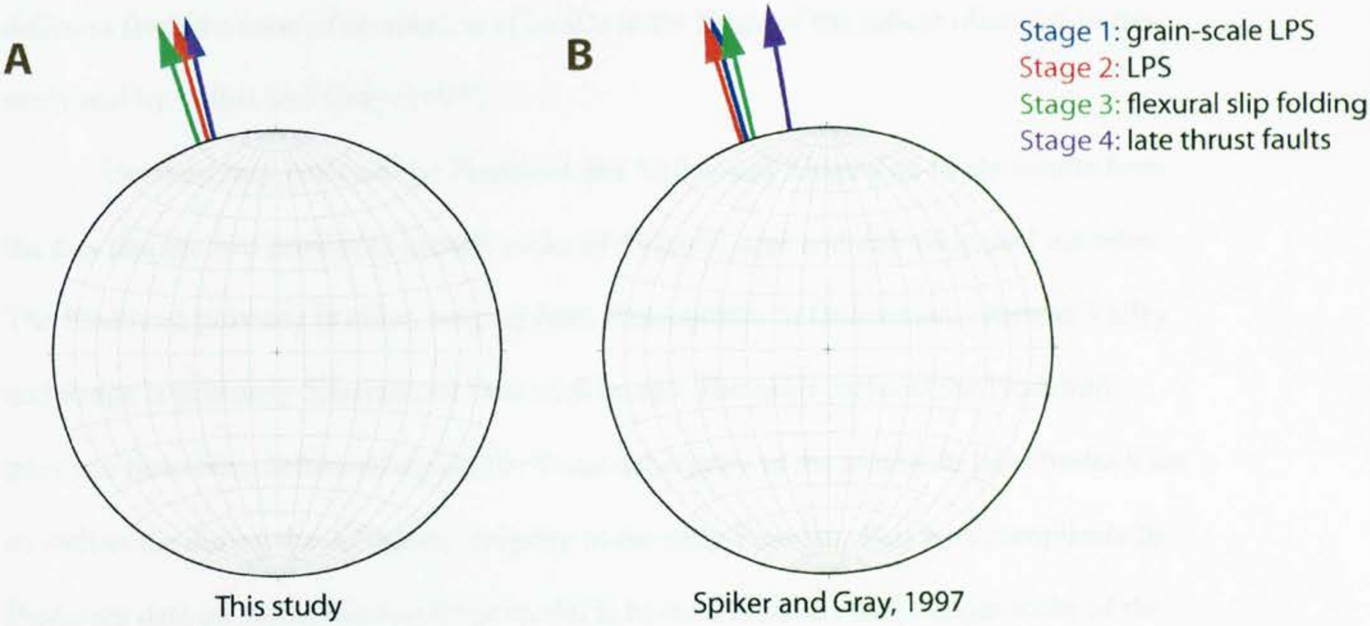


Figure 19: Equal area lower hemisphere stereonets showing the MSDs of each progressive stage of deformation found by this study and by Spiker and Gray (1997). Both studies reveal no rotation of MSD throughout the deformation sequence.

different from the zone of no rotation of MSDs in the hinge of the salient observed in this study and by Spiker and Gray (1997).

Inconsistency between the Piedmont and Valley and Ridge data likely results from the fact that the two provinces contain rocks of different ages and deformational histories. The Piedmont province is older, ranging from Precambrain to Ordovician, whereas Valley and Ridge is primarily Silurian and Devonian in age. The older rocks of the Piedmont province have been deformed during the Taconic orogeny in the middle to late Orodovician as well as the during the Alleghany orogeny in the early Permian. This may complicate the Piedmont data on which the two-stage model is based. However, the younger rocks of the Valley and Ridge Province were not deposited until after the Taconic orogeny, making it a better suited region for collecting structural data of Alleghanian deformation. A second possible source of inaccuracy in the two-stage model is that it is based largely on data collected before 1979, but it was not until that year that Nickelsen published the "Bear Valley" sequence of progressive deformation which is recognized as characteristic of the Alleghany orogeny. Without prior knowledge of the Alleghanian sequence, CW/CCW rotation of MSDs would be more easily interpreted as separate tectonic events rather than one orogeny with a rotation of stress trajectories.

IMPLICATIONS

Distinction between the single- and two-stage models is useful in classifying the Pennsylvania salient along the gradient of orogenic curves from primary arcs, which form in a curved shape from the onset of deformation, to secondary arcs, which are originally linear mountain belts that have later been bent by another force. Because the Pennsylvania salient appears to have been produced by a single tectonic event, it is improbable that the salient is a secondary arc. However, rotation of MSDs with time implies that primary inheritance of curvature is also unrealistic. Most likely, the Pennsylvania salient is an intermediate between these primary and secondary endmembers and can be classified as a progressive arc, where cruvature is acquired continuously throughout the evolution of the Alleghany orogeny.

Inherent in any progressive arc model for the Pennsylvania salient is a need for extension tangential to the foreland side of the salient. Tangential stretching accommodates the extensional stress that is acquired in the foreland as curvature develops (Marshak, 1988). However, the Pennsylvania salient has a notable lack of tangential stretching features. Nickelsen (1979) noted a few small late-stage cross-grabens on the "whaleback" anticline of the Bear Valley mine and Faill (1981) reported a series of minor conjugate strike-slip faults just west of the Alleghany front, but these account for only a small fraction of the tangential stretching that would be required for progressive bending of the Pennsylvania salient (Wise, 2004). The overprinting deformations of the two-stage model of Wise (2004) circumvent this problem, but in a progressive model like the single stage model of Gray and Stamatakos (1997), tangential stretching must be accounted for. Difficult to observe micro-scale processes, such as grain-boundary sliding, may have diffused extensional strain (Gray and Stamatakos, 1997), and strike-slip faults common throughout the Pennsylvania salient (Nickelsen, 1996) may accommodate curvature of the salient without the formation of large-scale tangential extension structures (Gray and Stamatakos, 1997).

The lack of rotation of MSDs during the progressive deformation sequence through the hinge of the Pennsylvania salient in the Valley and Ridge province has implications for the construction of balanced geologic cross-sections through the mountain belt. Such cross-sections are used to constrain the magnitude of shortening and orogenic mass balance. They require extensive field mapping and must meet several general requirements. A balanced cross-section must be both admissible (the cross-section agrees with all structures observed in the field) and viable (it can be returned to an undeformed state). Viability is not possible if any motion occurs outside the plane of the cross-section (Elliott, 1982), for instance, if there are multiple directions of shortening. Because a single-stage model with only one maximum shortening direction appears to be in effect through the hinge of the Pennsylvania salient, this is a suitable location through which to draw a balanced cross-section oriented parallel to the direction of maximum shortening.

CONCLUSIONS

This field-based investigation of progressive deformation led to several important conclusions regarding the tectonic evolution of the Pennsylvania salient:

1. MSDs at the study site in the southern Valley and Ridge in the hinge of the Pennsylvania salient remain essentially constant throughout the entire deformation sequence. MSDs span less than 20°, ranging from ~335° to ~351° with no trend either CW or CCW. The average MSD for the entire sequence is 342°.These results parallel findings at the north end of the salient's hinge at the Appalachian structural front where the average MSD for the same deformation sequence is 341° (Spiker and Gray, 1997).
2. The hinge of the Pennsylvania salient appears to coincide with an axis of no rotation of MSDs, consistent with the single-stage model of tectonic shortening which predicts

a constant MSD of ~340° through the hinge of the salient (Gray and Stamatakos, 1997).

3. The two-stage model of tectonic shortening of Wise (2004) is not in agreement with the structural data seen in the Valley and Ridge province, as this model predicts two distinct MSDs in the hinge of the salient; the first at ~325° and the second at ~290°. Inconsistency between the Piedmont and Valley and Ridge data likely results from complication of the Piedmont province by Taconic deformation, as well as that the piedmont data was collected prior to 1979 before knowledge of the characteristic Bear Valley sequence of the Valley and Ridge province.
4. The Pennsylvania salient fits the description of a progressive arc where curvature of the mountain belt was acquired continuously throughout the progression of the Alleghany orogeny. However, a progressive model must explain the lack of observed tangential stretching features in the Pennsylvania salient.
5. The hinge of the Pennsylvania salient is a reasonable transect through which to draw a geologic cross section of the Central Appalachians, as all deformation occurs along a constant tectonic shortening direction of ~340°.

REFERENCES

Boyer, S.E., 1995, Sedimentary basin taper as a factor controlling the geometry and advance of thrust sheets: American Journal of Science, v. 295, p. 1220-1254. Davis, D.M., and Engelder, T., 1985, The role of salt in fold-and-thrust belts: Tectonophysics, v. 119, p. 67-88.

Elliott, D., 1982, The construction of balanced cross-sections: Journal of Structural Geology, v.5, p.101.

Erslev, E.A., 1988, Normalized center-to-center strain analysis of packed aggregates: Journal of Structural Geology, v. 10, p. 201-209.

Evans, M.A., 1994, Joints and décollement zones in Middle Devonian shales: evidence for multiple deformation events in the central Appalachian Plateau: Geological Society of America Bulletin, v.106, p.447-460.

Faill, R.T., 1985, The Acadian orogeny and the Catskill delta, in Woodrow, D.L. and Sevon, W.D., eds, The Catskill delta: Geological Society of America Special Paper 201, p. 15-37.

Faill, R.T., 1997a, A geologic history of the north-central Appalachians, part 1: orogenesis from the Mesoproterozoic through the Taconic orogeny: American Journal of Science, v. 297, p. 551-619.

Faill, R.T., 1997b, A geologic history of the north-central Appalachians, part 2: the Appalachian basin from the Silurian through the Carboniferous: American Journal of Science, v. 297, p. 729-761.

Faill, R.T., 1998, A geologic history of the north-central Appalachians, part 3: the Alleghany orogeny: American Journal of Science, v. 298, p. 131-179.

Fry, N., 1979, Random point distributions and strain measurement in rocks: Tectonophysics, v. 60, p. 806-807.

Geiser, P.A., and Engelder, T., 1983, The distribution of layer parallel shortening fabrics in the Appalachian foreland of New York and Pennsylvania: Evidence for two non-coaxial phases of the Alleghanian orogeny, *in* Hatcher, R.D., Jr. et al., eds., Contributions to the tectonics and geophysics of mountain chains: Geological Society of America Memoir 158, p.161-175.

Gibb, R.A., 1978, Slave-Churchill collision tectonics: Nature, v. 271, p. 50-52.

Gibb, R.A., 1983, Model for suturing of Superior and Churchill plates: An example of double indentation tectonics: Geology, v. 11, p. 413-417.

Gray, M.B., and Mitra, G., 1993, Migration of deformation fronts during progressive deformation: evidence from detailed studies in the Pennsylvania Southern Anthracite region, U.S.A.: Journal of Structural Geology, v.15, p.435-450.

Gray, M.B., and Stamatakos, J., 1997, New model for evolution of fold and thrust belt curvature based on integrated structural and paleomagnetic results from the Pennsylvania salient: Geology, v. 25, p.1067-1070.

Hatcher, R.D., Jr., 1989, Tectonic synthesis of the U.S. Appalachians, *in* Hatcher, R.D., Jr., Thomas, W.A., and Viele, G.W., eds., The Appalachian-Ouachita orogen in the United States: Boulder, Colorado, Geological Society of America, Geology of North America, v. F2, p. 511-536.

Laubscher, H.P., 1972, Some overall aspects of the Jura dynamics: American Journal of Science, v. 272, p. 293-304.

Lawton, T.F., Boyer, S.E., and Schmitt, J.G.,1994, Influence of inherited taper on structural variability and conglomerate distributrion, Cordilleran fold and thrust belt, western United States: Geology, v. 22, p. 339-342.

Macedo, J., and Marshak, S., 1999, Controls on the geometry of fold-thrust belt salients: Geological Society of America Bulletin, v. 111, p. 1808–1822.

Markley, M., and Wojtal, S., 1996, Mesoscopic structure, strain, and volume loss in folded cover strata, Valley and Ridge province, Maryland: American Journal of Science, v.296, p.1674-1684.

Marshak, S., and Wilkerson, M.S., 1992, Effect of overburden thickness on thrust belt geometry and development: Tectonics, v. 11, p. 560-566.

Marshak, S., 2004, Salients, recesses, arcs, oroclines, and syntaxes; a review of ideas concerning the formation of map-view curves in fold-thrust belts, *in* McClay, K.R., ed., Thrust Tectonics and Hydrocarbon Systems: American Association of Petroleum Geologists Memoir 82, p. 131–156.

Mitra, S., 2002, Fold-accomodation faults: AAPG Bulletin, v. 86, p. 671-693.

Nickelsen, R.P., and Hough, V.D., 1967, Jointing in the Appalachian Plateau of Pennsylvania: Geological Society of America Bulletin, v. 4, p. 609-630.

Nickelsen, R.P., 1979, Sequence of structural stages of the Allegheny Orogeny, at the Bear Valley Strip Mine, Shamokin, PA: American Journal of Science, v. 279, p.225-271.

Nickelsen, R.P., 1988, Structural evolution of folded thrusts and duplexes on a first-order anticlinorium in the Valley and Ridge province of Pennsylvania, *in* Mitra, G., and Wojtal, S., eds., Geometries and mechanisms of thrusting, with special reference to the Appalachians: Geological Society of America Special Paper 222, p.89-106.

Nickelsen, R.P., 1996, Alleghanian sequential deformation on the SW limb of the

Pennsylvania salient in Fulton and Franklin Counties, south-central Pennsylvania: Pennsylvania Geological Survey, 61st Annual Field Conference of Pennsylvania Geologists Guidebook, 108 p.

Nickelsen, R.P., 2009, Overprinted strike-slip deformation in the southern Valley and Ridge *in* Pennsylvania: Journal of Structural Geology, v. 31, p. 865-873.

Pavoni, N., 1986, Regularities in the pattern of major fault zones of the Earth and the origin of arcs, *in* Wezel, F.C., ed., The origin of arcs: Amesterdam, Elsevier, p. 63-78.

Pennsylvania Geology Survey, 2007, Geologic map of Pennsylvania (color). 3rd ed., Scale 1:2,000,000.

Ramsay, J.G., 1967, Folding and Fracturing of Rocks: McGraw-Hill, NY. 568 p.

Rogers, H.D., 1858, The geology of Pennsylvania: Philadelphia, Pennsylvania, J.B. Liponcott, 1045 p.

Rogers, J., 1949, The tectonics of the Appalachians: New York, Wiley-Interscience, 271 p.

Schwartz, S.Y., and Van der Voo, R., 1983, Paleomagnetic evaluation of the orocline hypothesis in the central and southern Appalachians: Geophysical Research Letters, v. 10, p. 505–508.

Spiker, E.C., and Gray, M.B., 1997, A study of progressive deformation at the Alleghany front, Lycoming County, Pennsylvania.

Stamatakos, J., and Hirt, A.M., 1993, Paleomagnetic considerations of the development of the Pennsylvania salient in the central Appalachians: Tectonophysics, v. 231, p. 237-255.

Stamatakos, J.A., Hirt, A.M., and Lowrie, W., 1996, The age and timing of folding in the

Central Appalachians from paleomagnetic results: Geological Society of America Bulletin, v. 108, p. 815–829.

Tapponnier, P., and Molnar, P., 1976, Slip-line field theory and large-scale continental tectonics: Nature, v. 264, p. 319-324.

Thomas, W.A., 1977, Evolution of Appalachian-Ouachita salients and recesses from reentrants and promontories in the continental margin: American Journal of Science, v. 277, p. 511-522.

Thomas, W.A., 1983, Continental margins, orogenic belts, and intracratonic structures: Geology, v. 11. p. 270-272.

U.S. Geological Survey Open-File Report 2009-1042

Van der Pluijm, B.A., and Marshak, S., 2004, Earth Structure, 2nd edition: New York, W. W. Norton & Co., 656 p.

Weil, A.B., and Sussman, A.J., 2004, Classifying curved orogens based on timing relationships between structural development and vertical-axis rotations, in Sussman, A.J., and Weil, A.B., eds., Orogenic Curvature: Integrating Paleomagnetic and Structural Analysis: Geological Society of America Special Paper 383, p. 1–15.

Witlschko, D.V. and Chapple, W.M., 1977, Flow of weak rocks in the Appalachian Plateau folds: American Association of Petroleum Geologists Bulletin, v. 61, p. 653-670.

Wise, D.U., 2004, Pennsylvania salient of the Appalachians: a two-azimuth transport model based on new compilations of Piedmont data: Geology, v.32, p.777-780.

Wise, D.U., and Werner, M.L., 2004, Pennsylvania salient of the Appalachians: A two- stage model for Alleghanian motion based on new compilations of Piedmont data, *in* Sussman, A.J., and Weil, A.B., eds., Orogenic Curvature: Integrating Paleomagnetic and Structural Analyses: Geologic Society of America Special Paper 383, p.109-120.

Younes, A., and Engelder, T., 1999, Fringe cracks: A key data set for the interpretation of progressive Alleghanian deformation of the Appalachian Plateau: Geological Society of America Bulletin, v.111, p.219-239.

Yonkee, A., and Weil, A.B., 2009, Reconstructing the kinematic evolution of curved mountain belts: internal strain patterns in the Wyoming salient, Sevier thrust belt, U.S.A.: Geological Society of America Bulletin, v. 122, p. 24-49.

Zhou, M., and Jacobi, R.D., 1997, Formation of regional cross-fold joints in the northern Appalachian Plateau: Journal of Structural Geology, v.19, p.817-834.