Unraveling the central Appalachian fold-thrust belt, Pennsylvania: The power of sequentially restored balanced cross sections for a blind fold-thrust belt

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ABSTRACT

We present a kinematic model for the sequential development of the Appalachian fold-thrust belt (eastern U.S.) across a classic transect through the Pennsylvania salient. New map and strain data are used to create a balanced geologic cross section from the southern edge of the Valley and Ridge Province to the northern Appalachian Plateau. This region of the central Appalachian fold-thrust belt is an ideal location to illustrate the incorporation of strain data in balanced cross sections, because it cannot be balanced without quantifying grain-scale strain. We use a sequentially restored, balanced cross section to show how layer-parallel shortening (LPS) is distributed above and ahead of thrust and fold shortening and constrain the geometric and kinematic evolution of a passive roof duplex. By combining line length and area balancing of a kinematically viable cross section with LPS estimates in both the Valley and Ridge Province (20%) and Appalachian Plateau (13%), we document the total magnitude of shortening in both the folded cover sequence and the duplexed lower layer of the fold-thrust belt. Restoration of the cross section indicates a total of 77 km (22%) of shortening between the southern margin of the Valley and Ridge Province in central Pennsylvania and a pin line immediately north of the northern limit of documented LPS in the foreland. The 24 km (13%) of LPS on the Appalachian Plateau is interpreted as being above the Salina (salt) décollement. This magnitude of shortening is 14 km greater than the amount of displacement on the Nittany Anticlinorium, the northernmost structure of the fold-thrust belt that cuts upsection from the Cambrian Waynesboro Formation to the Silurian Salina décollement. Because the fault that cores the Nittany Anticlinorium can only facilitate 10 km of shortening on the plateau, an early history of Appalachian Plateau LPS in Silurian and younger rocks is required to balance the section. We propose that the additional 14 km of LPS on the plateau occurred early in the deformation history and was kinematically linked to two fault-bend folds that have a lower décollement in the Cambrian Waynesboro Formation and an upper, subhorizontal detachment in the Silurian Wills Creek Formation (in the Valley and Ridge) and the Salina Group on the Appalachian Plateau. This upper detachment feeds displacement from these early horses in the duplex system onto the Appalachian Plateau and is expressed there as LPS shortening. This early shortening is followed by the development of in-sequence horses that repeat the mainly thrust-faulted Cambrian–Ordovician sequence using both the main décollement in the Cambrian Waynesboro and the Ordovician Reedsville Formations as an upper detachment horizon. In the south, shortening in the Late Ordovician through Devonian layers is accommodated by both LPS and forced folding of the overlying folded cover sequence. We propose that the Reedsville Formation becomes weaker to the north, facilitating shorter wavelength detachment folds. The development of gentle open folds on the Appalachian Plateau, as well as the last 10 km of LPS on the plateau, is linked to the most forelandward horse in the duplex. This horse forms the broad Nittany Anticlinorium, the northern boundary of the Valley and Ridge.

INTRODUCTION

The northern section of the central Appalachian fold-thrust belt (eastern U.S.) is a classic example of a blind thrust system. At its northernmost end, the fold-thrust belt sweeps eastward, creating the broad arc of the Pennsylvania salient (Fig. 1). Although previous research in the central Appalachians has made considerable progress toward quantifying how shortening is distributed among microscopic (e.g., Smart et al., 1997; Thorbjornsen and Dunne, 1997), mesoscopic (e.g., Smart et al., 1997; Hogan and Dunne, 2001), and map-scale structures (e.g., Herman, 1984; Hatcher, 1989; Mitra, 2002), a fully balanced section across the Valley and Ridge, through the Pennsylvania salient, where slip from deeper structures is tracked to structures that accommodate shortening in the upper layers to surface, has yet to be constructed. The first cross sections highlighted significant discrepancies between the amount of shortening that can be documented in the folded sequence of Ordovician–Pennsylvanian strata and the amount of shortening needed in the imbricated sequence of Cambrian–Ordovician carbonates to fill space between the overlying folds and the seismically imaged basement (Gwinn, 1970; Herman, 1984; Herman and Geiser, 1985). Significant layer-parallel shortening (LPS) has occurred throughout the Pennsylvania salient (i.e., Nickelsen, 1966, 1979; Engelder, 1979a; Gray and Mitra, 1993), and balanced sections must take this shortening into account as well as other mechanisms of strain such as submap-scale mesostructures including joints, faults, and fold arrays.

Some of the earlier attempts at constructing cross sections invoked LPS to account for the proposed 72 km discrepancy in the restored lengths of the imbricated carbonate sequence and mainly folded strata (Fig. 2). These solutions require 28% LPS in the folded cover strata across the Valley and Ridge Province (Herman, 1984; Hatcher, 1989). However, the 28% LPS was not directly measured; rather, this is the magnitude necessary to reconcile differences in shortening between the proposed imbricated carbonate sequence and the observed folded

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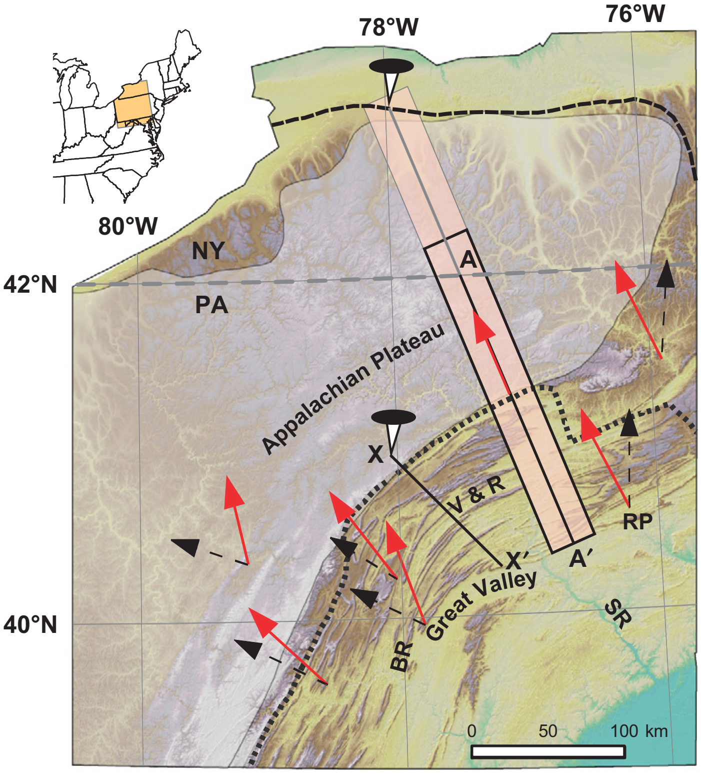


Figure 1. Shaded relief map of the Pennsylvania salient and surrounding region showing the subsurface extent of Silurian salt horizons (light blue shaded region) (modified after Davis and Engelder, 1985); northern limit of layer-parallel shortening (dashed line, after Geiser and Engelder, 1983); position of the Alleghany front (dotted line); compilation of tectonic transport vectors (red arrows—early transport direction; black arrows—late transport direction) from Gray and Stamatakos (1997). Shaded red box—study area; heavy black outlined box—extent of Figure 4; bold line labeled X–X′—line of section shown in Figure 2; bold line labeled A–A′—line of section shown in Plate 1. BR—Blue Ridge; RP—Reading Prong; SR—Susquehanna River; V&R—Valley and Ridge; NY, PA—New York, Pennsylvania. Note that the line of section parallels the transport direction along the axis of no rotation of the maximum shortening direction.

cover strata in early cross sections (Herman, 1984; Hatcher, 1989).

In this study we pin the cross section in the undeformed foreland and treat LPS through the Valley and Ridge and LPS translation across the Appalachian Plateau as intrinsically linked to the thrust faults interpreted to underlie the folds. LPS is quantified through compilations of existing (Engelder, 1979a; Nickelsen, 1963, 1983; Faill and Nickelsen, 1999) and new finite strain analyses along the profile. By placing the pin line beyond the limit of documented deformation (Engelder, 1979b; Geiser and Engelder, 1983), both LPS and translation of strain across the Appalachian Plateau are included in the balanced cross section. In addition, we use sequential restoration to test that slip on deeper structures is fed through linked fault systems to the slip on shallower structures and then eventually to the surface. We present a section that is both line-length and area balanced, and show that fault slip is conserved along the entire path of a thrust system, ensuring viability (Boyer and Elliott, 1982; Woodward et al., 1989; McQuarrie, 2002; McQuarrie et al., 2008; Robinson, 2008).

GEOLOGIC BACKGROUND

The structures in the Valley and Ridge are the result of tectonic shortening and thickening associated with the closure of the Iapetus Ocean and culminating in the Permian continent-continent collision of Gondwana with Laurentia in the Alleghanian orogeny (i.e., Rodgers, 1949; Hatcher, 1989; Stamatakos et al., 1996; Faill, 1998). The arc of the Pennsylvania salient links two relatively linear segments, the north-south–trending Blue Ridge to the southwest and the east-west–trending Reading Prong to the northeast (Fig. 1). The shape and position

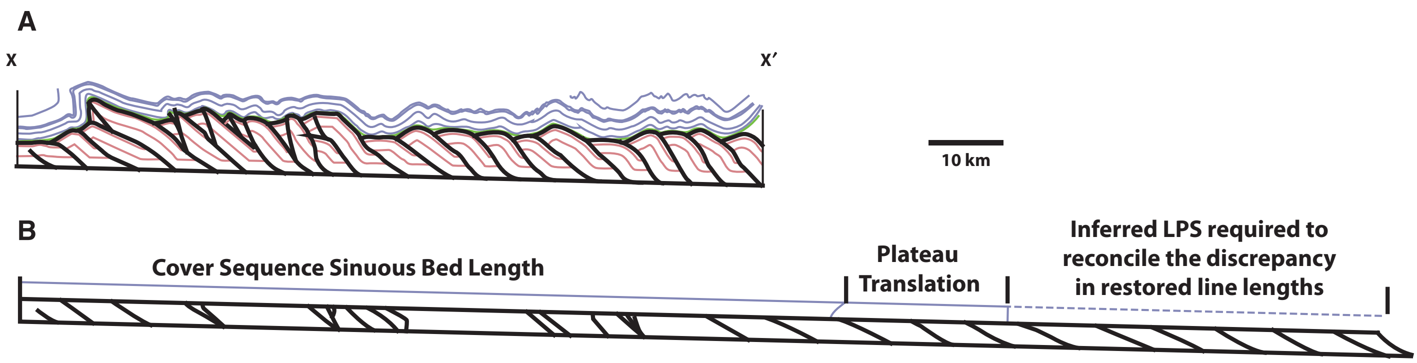


Figure 2. (A) Simplified geologic cross section extending from the southern boundary of the Valley and Ridge physiographic province to the pin line on the Appalachian Plateau. (B) Restored cross section from X–X′ highlighting the 28% discrepancy in the restored line lengths between imbricated carbonate sequence and overlying mainly folded strata. Cross section is simplified from Herman (1984). LPS—layer-parallel shortening.

of the Pennsylvania salient have been attributed to the tectonic inheritance of the Iapetan rifted margin of eastern Laurentia (Beardsley and Cable, 1983; Thomas, 1977, 2006; Ong et al., 2007). In the Susquehanna River valley, along the line of the section, the ridges of the Valley and Ridge Province trend ~070°. Here the Valley and Ridge is defined as an ~110-km-wide swath of alternating valleys and ridges with moderate (<400 m) relief.

The Paleozoic strata exposed at the surface through the central Appalachian Valley and Ridge Province are part of an unmetamorphosed, low-temperature (<300 °C) foreland basin sequence (Fig. 3). The structure of the Valley and Ridge Province of the Alleghanian fold-thrust belt is three tiered: an uninvolved Neoproterozoic crystalline and sedimentary rock basement, a faulted sequence of Cambrian–Ordovician carbonates and flysch, and a mainly folded cover sequence of Ordovician–Pennsylvanian foreland basin siliciclastic rocks that is decoupled from underlying strata along a passive roof-thrust detachment (Boyer and Elliot, 1982; Gwinn, 1964, 1970; Herman, 1984; Herman and Geiser, 1985; Onasch and Dunne, 1993; Perry, 1978; Scanlin and Engelder, 2003a). The fold-thrust belt is separated from the basement by a regional décollement in the middle Cambrian Waynesboro Formation (Gwinn, 1964, 1970; Rodgers, 1963, 1970) (Fig. 3). Above the basal décollement, imbricate thrusts are presumed to ramp upward from the lower décollement horizon (Herman, 1984; Geiser, 1988a; Faill, 1998) and coalesce with a roof thrust within the Ordovician Reedsville Formation (Fig. 3). A very limited number of faults has been recognized in the folded section above the roof thrust. These faults have small offsets identified at the surface, or are interpreted as being blind. It is suggested that the blind faults merge upsection into local décollements that are oriented subparallel to bedding and manifest as narrow (<30 cm thick) zones of strain localization (i.e., Nickelsen, 1986, 1988).

Shortening in the Ordovician–Pennsylvanian rocks is expressed at the surface as folds with 50–60-km-long hinges and wavelengths of ~7–12 km that are seen in the first-order topography of the Valley and Ridge (Fig. 1). These folds are characterized by narrow hinges relative to their wavelength. The long, continuous anticlines yield aspect ratios (half wavelength to axial length ratio) of 10:1–16:1, outside the norm of 5:1–10:1 common for buckle folds (i.e., Sattarzadeh et al., 2000). This suggests that the first-order folds of the Valley and Ridge are forced folds, controlled by fault-bend folding in subsurface layers (Sattarzadeh et al., 2000). Thus, the geometry of the horses within the

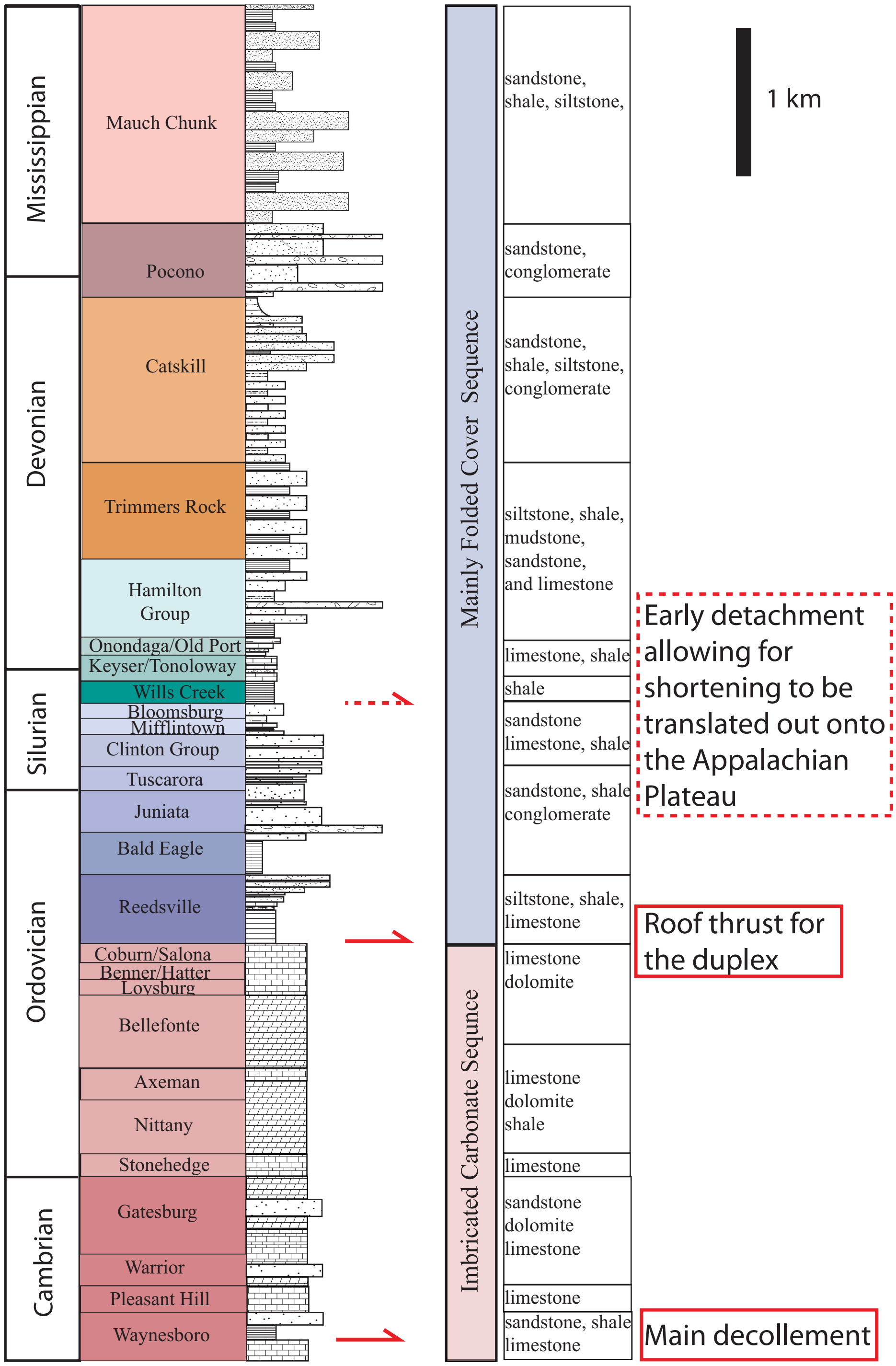


Figure 3. Stratigraphic column for the central Appalachian Valley and Ridge Province (compiled from Ayrton, 1963; Colton, 1963; Hoskins et al., 1963; Sutton, 1963; Gwinn, 1964; Dyson, 1967; Faill and Wells, 1974; Hoskins, 1976; Faill and Nickelsen, 1973, 1977, 1978; Berg and Edmunds, 1979; Wells and Bucek, 1980; Inners, 1981, 1997; Berg et al., 1983, 1993; Epstein, 1986; Edmunds, 1993, 1996). Note that the Ordovician Martinsburg Formation is the Great Valley temporal equivalent to the Reedsville Formation in the Valley and Ridge.

imbricated carbonate sequence places a first-order control on the fold train in overlying cover strata (i.e., Gwinn, 1964; Shumaker et al., 1985; Ferrill and Dunne, 1989; Meyer and Dunne, 1990; Wilson and Shumaker, 1992). The folds in the Valley and Ridge are asymmetric, with steeper northern limbs (Rodgers, 1949).

The northern boundary of the Valley and Ridge structural province is defined by the Alleghany front (Figs. 1 and 4), where an abrupt change in the geometry of folds at the surface and the geometry of the basal décollement occurs. At the Alleghany front, the cover sequence is folded into moderately tight northward-inclined asymmetric folds that yield to broad, gentle (10 km wavelength, 100 m amplitude) folds with gently dipping (<2°) limbs (Wedel, 1932) characteristic of the Appalachian Plateau (Fig. 4). The change in folding style is coincident with the subsurface extent of Silurian salts (Fig. 1) (Rodgers, 1963; Prucha, 1968; Wiltschko and Chapple, 1977; Davis and Engelder, 1985) and occurs just to the north of a seismically imaged ramp in the basal décollement from the Cambrian Waynesboro Formation (beneath the Valley and Ridge) to the Wills Creek–Salina interval in the Late Silurian (beneath the Appalachian Plateau; e.g., Gwinn, 1964; Beardsley et al., 1999; Scanlin and Engelder, 2003a, 2003b).

The dearth of map-scale faults has significant implications for estimating the magnitude of shortening within the cover section as well as the geometry of shortening in the fold-thrust belt as a whole. Previous restorations of both the folded cover rocks and the underlying imbricated carbonate sequence indicate 28% less shortening in the folded strata than in the underlying faulted strata (Herman, 1984; Geiser, 1988b). Consequently, significant mesoscale and microscale mechanisms of shortening are invoked (i.e., LPS) to shorten the cover sequence and balance the deformation throughout the section.

Intergranular twin and translation gliding, grain-boundary sliding, intergranular cataclastic flow, and/or crenulation (grain rotation) and dissolution are recognized as significant mechanisms of LPS within the folded cover sequence (i.e., Faill and Nickelsen, 1973, 1999; Nickelsen, 1972, 1986; Groshong, 1975; Engelder, 1979b). Penetrative deformation manifested as distorted mud-crack polygons, reduction spots, and fossils occurred in two phases. In the first phase, bed-parallel stylolites in carbonate layers and distorted fossils and reduction spots in the clastic layers record early pre-Alleghanian compaction perpendicular to bedding. Reduction spots are compacted 20%–30% into oblate ellipsoids (Nickelsen, 1983; Faill and Nickelsen, 1999). The second phase of penetrative deformation at the beginning of the Alleghanian folding consists of LPS. This second phase is characterized by shortening oriented perpendicular to fold axes. Measurements of crinoid ossicles in the bedding surface yield an estimated range of ellipticity (R_s) values of 1.05–1.28 with a mean of 1.18 ± 0.07, with the short axis of the ellipse perpendicular to the regional fold axes in the Valley and Ridge (Nickelsen, 1983) and parallel to original bedding, defining LPS. These values of LPS are generally greater than values measured beyond the northern limit of Alleghanian folding on the Appalachian Plateau (i.e., Nickelsen, 1966; Engelder and Engelder, 1977; Geiser and Engelder, 1983). Discrete structural stages have been identified through the Pennsylvania Valley and Ridge fold-thrust belt based on the identification and relative ages of microscale to macroscale structures (e.g., Gray and Mitra, 1993). The structural stages involve, in order, LPS, top-to-the-north shear, main folding, and fold modification by low-angle thrust faulting (Nickelsen, 1979; Gray and Mitra, 1993; Faill and Nickelsen, 1999).

Because deformation extends north of the Alleghany front across the southern Appalachian Plateau, we pin the cross section in New York State north of where Wedel (1932) documented gently folded strata with dip magnitudes rarely exceeding 2° (Figs. 1 and 5). The pin line is located north of the position where calcite twin data from Devonian shales document ≤2% of finite strain (Engelder, 1979b) (Fig. 5). The southern end of the cross section is placed along the southern margin of the Valley and Ridge, within the Great Valley. We argue that this southern boundary provides a natural break that coincides with a change in both deformation style and history. The Great Valley coincides with exposed Ordovician Martinsburg Shale at the core of a structural high and is in close proximity to the westernmost fault of the Triassic basin extensional system (Berg et al., 1980). Down-plunge projections to the north and south suggest that the Martinsburg structural high is cored by culminations in basement and Cambrian rocks that are structurally lower than the Cambrian Waynesboro décollement through the Valley and Ridge (Berg et al., 1980). Thus, shortening due to duplication of these deeper rocks would balance shortening in the Valley and Ridge, not add to it. In addition, the Martinsburg Formation is dominated by extensive pressure solution cleavages recording ≥50% LPS in shale, much of which predates the Alleghanian orogeny (Wright and Platt, 1982; Ganis and Wise, 2008; Wise and Ganis, 2009).

Systematic variations in shortening directions are recognized between the early and late stages of Alleghanian deformation across the Valley and Ridge and Appalachian Plateau to the north. East of the Susquehanna River valley studies reveal 25°–30° of clockwise rotation in the orientation of Alleghanian shortening direction as a function of time based upon the progressive deformation sequence (Fig. 1) (Nickelsen, 1979; Geiser and Engelder, 1983; Gray and Mitra, 1993; Zhao and Jacobi, 1997; Younes and Engelder, 1999). In contrast, farther to the west in the Blue Ridge segment of the Valley and Ridge and adjacent parts of the Appalachian Plateau and Great Valley, studies reveal 15°–45° of counterclockwise rotation in shortening direction (Fig. 1) (Nickelsen, 1988, 2009; Evans, 1994; Markley and Wojtal, 1996). Early-stage LPS indicates shortening directions on both limbs of the salient that are subparallel, trending ~320°–350°. In addition to linking the east-northeast–trending Reading Prong to the south-southwest–trending Blue Ridge segments of the Appalachians, the hinge of the Pennsylvania salient coincides with the axis of no rotation in the shortening direction over time (Fig. 1) (Spiker and Gray, 1997; Gray and Stamatakos, 1997).

METHODS

In this study, geologic mapping at a scale of 1:100,000 and finite strain analysis are combined to constrain a northwest-trending geologic cross section across the Pennsylvania salient segment of the Valley and Ridge and Appalachian Plateau. Specifics of each of the methods are described in the following.

Geologic Mapping

Our geologic map and cross section are based upon a compilation of previous mapping (Wedel, 1932; Boyer, 1972; Hoskins, 1976; Faill and Wells, 1974; Faill et al., 1977; Faill, 1979; Wells and Bucek, 1980; Inners, 1997). These data are combined with new structural measurements collected along the transect (Figs. 1 and 4). Original maps were made at scales of 1:6,000–1:24,000 and compiled on a 1:100,000 topographic base. The selected transect is oriented 337°, parallel to the axis of no rotation of the shortening direction (Spiker and Gray, 1997; Gray and Stamatakos, 1997).

Strain Measurements

For finite strain measurements, we targeted distorted crinoid ossicles in siltstone lithologies and distorted grains (fine to coarse grain size) in nonfossiliferous lithologies throughout the exposed stratigraphic section to test the lithologic control on LPS values. LPS finite strain

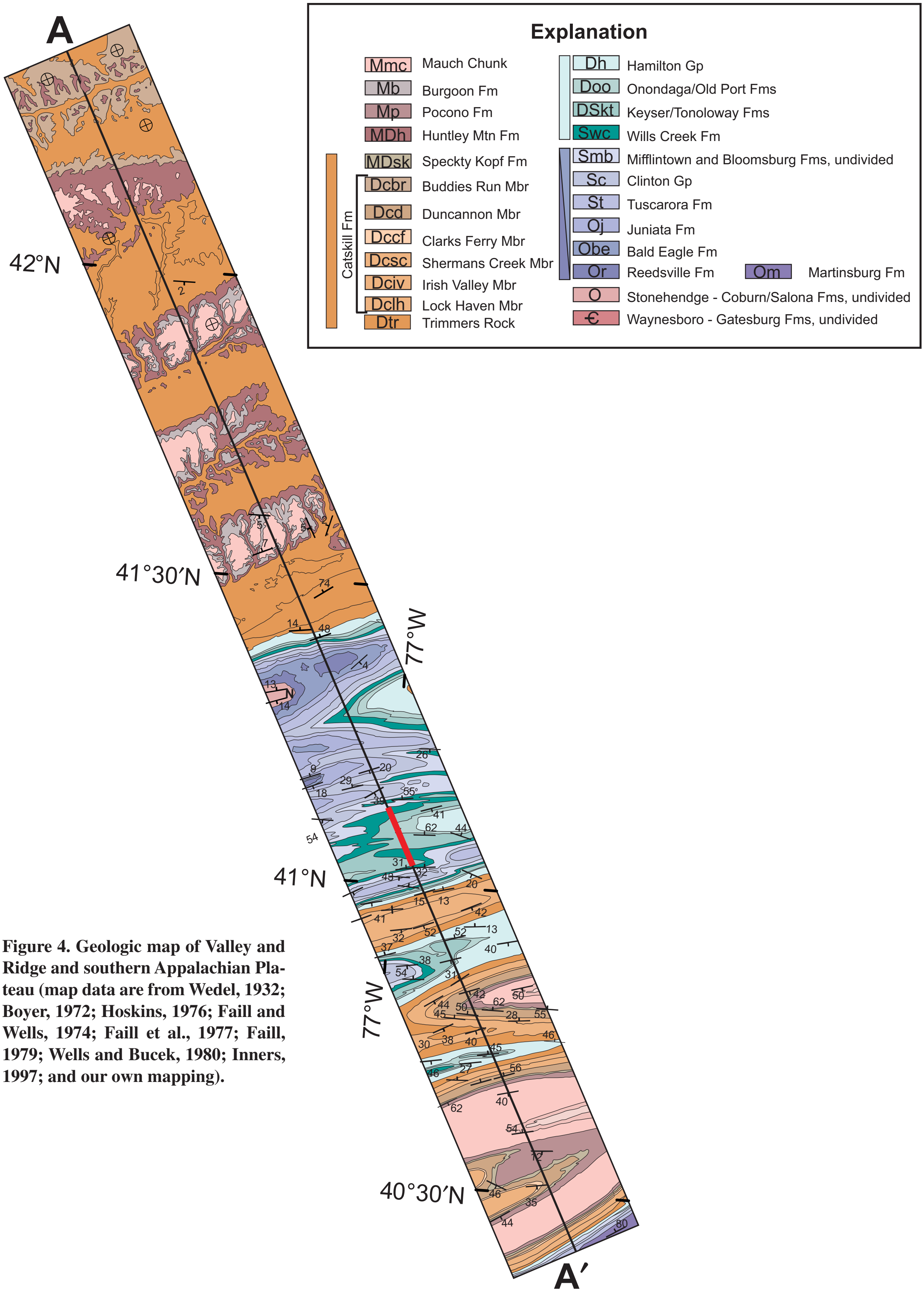


Figure 4. Geologic map of Valley and Ridge and southern Appalachian Plateau (map data are from Wedel, 1932; Boyer, 1972; Hoskins, 1976; Faill and Wells, 1974; Faill et al., 1977; Faill, 1979; Wells and Bucek, 1980; Inners, 1997; and our own mapping).

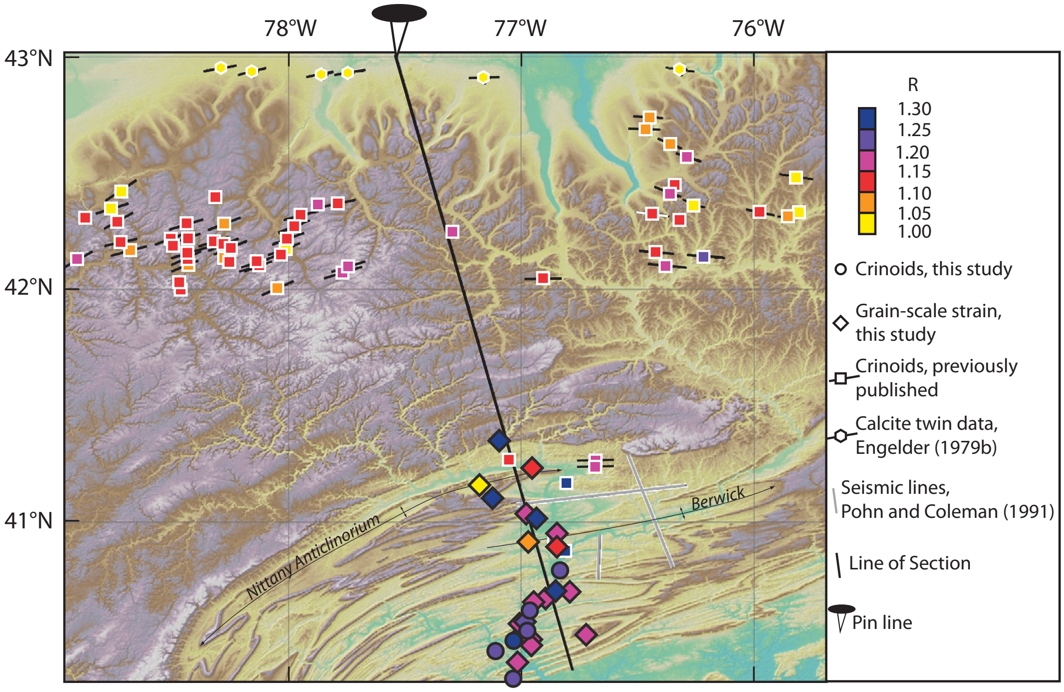


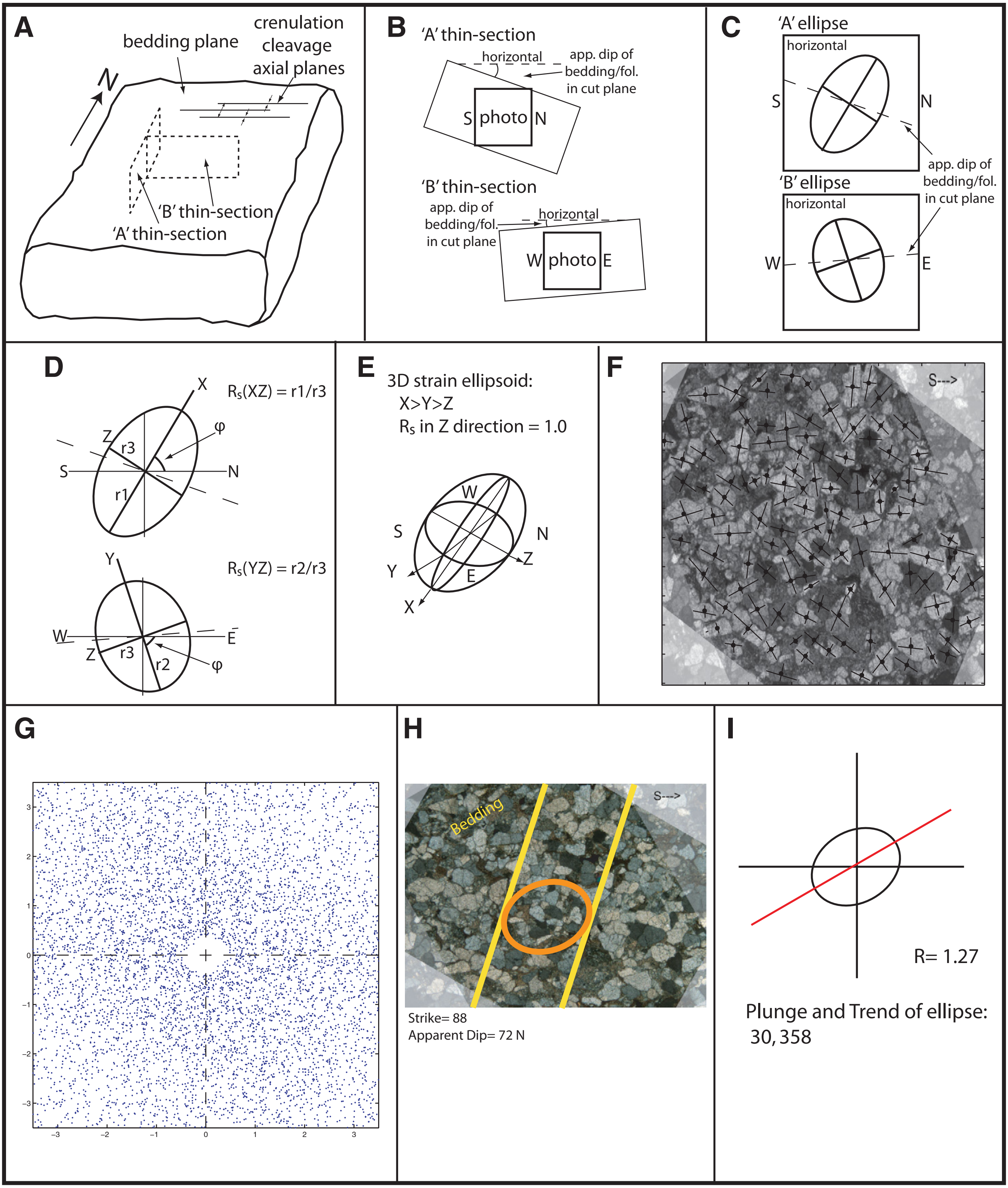
Figure 5. Shaded relief map of the study area and surrounding regions showing the distribution and magnitude of measured finite strain (from Engelder, 1979b; Slaughter, 1982; Geiser and Engelder, 1983; Nickelsen, 1983; and this study [larger symbols with black outlines]), locations of published seismic lines (Pohn and Coleman, 1991), and the line of cross section in Plate 1.

is most readily observed and quantified by measuring distorted crinoid ossicles deposited on bedding planes. Deformed crinoid ossicles are used to constrain bed-parallel strain at five previously unreported sites exposed across the Valley and Ridge Province in siltstones of the Devonian Trimmers Rock Formation. Strain measurements were made on oriented bedding-plane surfaces by measuring the long and short axes of individual crinoid ossicles (10–30 per sample) and the strike of the long axis on a weathered bedding plane. Following the procedure outlined by Engelder and Engelder (1977), measurements were collected directly on the bedding plane. Individual ossicles were measured to the nearest 0.1 mm using a digital caliper, by a minimum of two people. The representative strain ellipse was determined from the data using an algebraic method for strain estimation (Shimamoto and Ikeda, 1976). Therefore, the axial ratio and orientation of the strain ellipse for each individual site are an average of ~40 strain ellipses obtained by more than one person.

The magnitude and orientation of finite strain within nonfossiliferous units of the cover sequence were analyzed from oriented samples using the normalized Fry (Erslev, 1988) method for finite strain analysis of quartz grains. The normalized Fry method is an improved version of the Fry method (Fry, 1979; Ramsay and Huber, 1983), and allows more precise determination of bulk strain by correcting for the effects of variable sorting and packing (Erslev, 1988). We collected 28 samples from a range of rock units with different grain sizes, different bed thicknesses, and from different units that exhibit different deformation mechanisms such as cleavage, pressure solution features, and wedge faulting. Two perpendicular thin sections were cut from each oriented sample, one normal to bedding and parallel to the transport direction (A cut) and the other normal to bedding but perpendicular to the transport direction (B cut). For several samples, a bedding-parallel cut was also made (C cut) (Figs. 6A–6C).

For a single analysis, grain center locations and the lengths of long and short axes of 150–200 closely packed quartz clasts were measured off of at least two sets of photomicrographs per slide. The lengths of the axes were used to normalize the distance between the grain centers (e.g., Erslev, 1988), allowing for the normalized plotted grain centers to define a ring of high-density points surrounding a vacancy field that illustrates the shape of the strain ellipse for the sample (Fig. 6). The ellipticity (R_s) value for the best-fit ellipse is the ratio of the long to short axes of the contact between the vacancy field and the ring of high point density. The angle of inclination, φ, of the long axis of the best-fit strain ellipse (Ramsay and Huber, 1983), is measured relative to a horizontal reference line (Figs. 6F–6I).

Figure 6 (*on following page*). Diagrams and photographs illustrating strain calculation techniques (see text). (A) Orientations of A and B thin-section cuts relative to bedding or crenulation cleavage axial planes. (B) Thin sections are rotated by the apparent dip angle of bedding to the orientation of the maximum shortening direction in the cut plane, so that horizontal on the photo represents horizontal in space. (C, D) Strain ellipsoids on A and B thin sections from normalized Fry analysis; R_s (XZ) is ratio of longest and shortest axes (r1/r3), and R_s (YZ) is ratio of middle and shortest axes (r2/r3). Angle between horizontal and long axis of ellipse is ϕ (down to north or south in A cuts; down to east or west in B cuts). (E) Three-dimensional strain ellipsoid resulting from combining two-dimensional A and B strain ellipses; X > Y > Z, R_s of Z = 1.0. (F, G) Examples of normalized Fry method; grain axial ratios and center positions of ≥150 quartz grains are input to make normalized Fry plot. The ratio of long and short axes of the ellipse defined by the shape of the outside edge of the vacancy field is used to obtain R_s value, and ϕ is the orientation of the long axis relative to horizontal. (H) Orientation of bedding (yellow lines) and bulk strain ellipse (as constrained by the normalized Fry method) draped on a photomicrograph of sample 10. (I) Simplified diagram showing the R_s values of sample 10.


A
crenulation cleavage axial planes
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'B' thin-section
app. dip of bedding/fol. in cut plane
horizontal
W photo E
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'A' ellipse
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app. dip of bedding/fol. in cut plane
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3D strain ellipsoid:
X>Y>Z
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30, 358

Figure 6.

Optically continuous overgrowths on detrital quartz grain boundaries have the potential of biasing grain-to-grain strain calculations (Houseknecht, 1988; Dunne et al., 1990). However, comparison of strain ratios we obtained from cathodoluminescence photomicrographs (which highlight the detrital quartz grain boundaries; e.g., Houseknecht, 1988) to those from optical photomicrographs indicated no variation in strain ratios. Because the continuous overgrowths on the detrital quartz grains had no effect on the quantification of finite strain in our samples, we use optically defined grain boundaries to constrain the finite strain.

Two-dimensional ellipses from "A", "B", and "C" thin sections for a subset of the samples were combined to determine the three-dimensional strain ellipsoid of each sample. The two-dimensional elliptical data from three sections were analyzed using the best-fit ellipsoid program developed by Mookerjee and Nickleach (2011) to determine the best-fit strain ellipsoid using the least squared approach.

For the "A" and "B" cuts of each strain sample, the shared axis is perpendicular to bedding. Almost all samples match a strain field where the Z axis is the transport direction, the Y axis is approximately strike parallel, and the X axis is perpendicular to bedding. The difference between Y and X R_s values vary between 0.01 and 0.18, with a bedding-perpendicular X direction in all but one sample. However, in practice, the natural variability in X and Y R_s values average out to essentially the same value (1.22 ± 0.05). Since the "A" cut is parallel to transport direction and contains the maximum and minimum strains, and we have obtained similar R_s values for ZX and ZY, we discuss the strain ellipsoid in terms of simple ratios of the long axis (X or Y) to Z with the Z axis being assigned an R_s (tectonic ellipticity) of 1.0 (Figs. 6D, 6E). The R_s values of the long axis of the ellipse are (R_{yz}) for crinoid samples and (R_{xz}) for Fry analyses on nonfossiliferous samples.

Balanced Cross Section and 2DMove Reconstruction

In order for a cross section to be balanced, it must be both admissible and viable. Embedded in the notion of viability is the assumption that little or no motion occurs in or out of the plane of section (Dahlstrom, 1969; Elliott, 1983; Woodward et al., 1989). Also inherent in viability is that the displacement path of each structure is known such that the structures can be restored to an unstrained state, and that fault slip is conserved through the entire fault system (Dahlstrom, 1969; Boyer and Elliott, 1982; Elliott, 1983; Geiser, 1988b; Woodward et al., 1989). In the case of blind thrusts and LPS, this requires that as the fault loses displacement in the direction of transport, equivalent amounts of shortening are taken up by folding or LPS.

The balanced cross section presented here was constructed using the sinuous bed method (Dahlstrom, 1969). Faill (1969, 1973) documented that most of the folds through the Valley and Ridge are flexural-slip kink folds with planar limbs and narrow hinges. Along our section, we observe both narrow anticlinal hinges adjacent to broad synclines as well as narrow syncline hinges adjacent to broad topped anticlines (Plate 1). Concentric folding was maintained in the Appalachian Plateau, as field observations indicate that folding in the plateau is concentric and the shallow dips observed in the plateau do not create significant space problems between the observations at the surface and the mobile salt that is accommodating the folding.

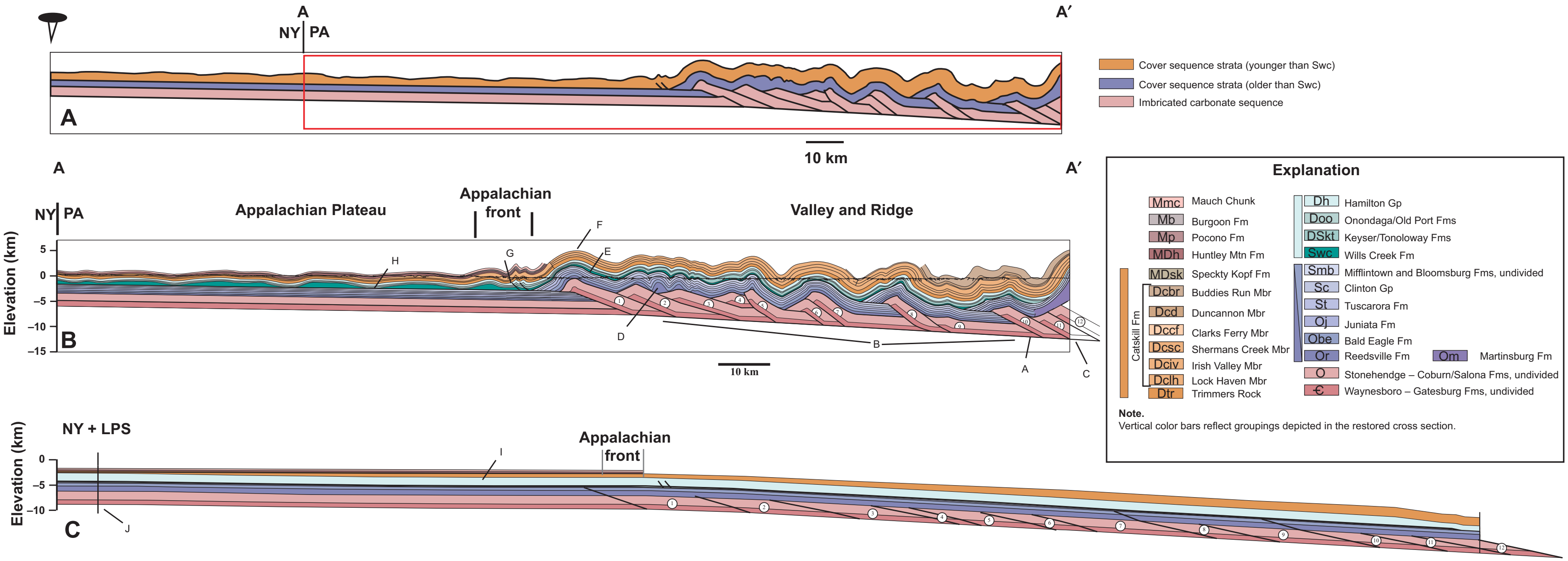
The pin line for the cross section is located 159 km north of the Alleghany front, where finite strain is 0 (Engelder, 1979b) (Fig. 5). The southern end of the cross section is placed along the southern margin of the Valley and Ridge, within the Great Valley (Fig. 1). Although the pin line for the cross section is beyond the northernmost extent of deformation, the gentle folds on the Appalachian Plateau end at the New York–Pennsylvanian border (Plate 1A). Thus, we use the state border as the northern edge of the detailed cross section (Plate 1B) and restored sections (Plate 1C) and show the 73 km of horizontal strata from the border to the pin line in Plate 1A. The original cross section and restored section were drawn and balanced by drafting both sections simultaneously by hand. The cross section was then digitized and imported into 2DMove (Midland Valley Exploration, Ltd.) to create the sequentially restored cross sections.

Forward modeling a cross section in 2DMove requires linking all folds to fault-parallel flow of material. Fault-parallel flow matches the first-order features of the fold-thrust belt, such as the forced fault-bend folds above horses in Cambrian–Ordovician strata, but cannot produce detachment folds. Detachment folding, such as on the Appalachian Plateau or in portions of the Valley and Ridge, cannot be forward modeled by 2DMove. Consequently, we use 2DMove to sequentially undeform the balanced cross section. We do this by first sequentially unfaulting the duplexed Cambrian–Ordovician section in order from north to south assuming the fold-thrust belt behaved as a simple forward-propagating system. We used the fault-parallel flow algorithm in 2DMove to remove faulting and restore many of the forced folds in the cover sequence. Because the displacement taken up by faulting is greater than that taken up by folding, this restoration also opens up gaps in the cover sequence that reflect (and require) shortening through LPS. Finally, we remove any remaining folding above restored horses using the "Unfold" algorithm in 2DMove. To calculate the shortening magnitude and percent shortening of the entire section, we add the 73 km length between the state line and the pin line in the undeformed foreland to account for the full section length.

QUANTIFICATION OF STRAIN

We derived 28 strain measurements from samples collected throughout the Valley and Ridge Province (Tables 1 and 2; Fig. 5); 27 samples were collected from exposures in the folded cover sequence and one sample was collected from the proposed unstrained faulted sequence. Ellipticities for the five strain ellipses measured using deformed crinoid ossicles on the bedding plane (R_{yz}) range from 1.19 ± 0.01 to 1.26 ± 0.01 with a mean of 1.23 ± 0.02 (Table 1). Like previously reported analyses, the strikes of the long axes of the bedding-parallel strain ellipses (the Y direction) are parallel to the strikes of folds (i.e., Faill, 1973; Nickelsen, 1983; Faill and Nickelsen, 1999), and the magnitudes of strain are consistent with those previously reported (Nickelsen, 1983) (Fig. 5). To document the strain across the Appalachian Plateau, we compiled strain estimates of deformed crinoid ossicles measured at 53 sites (Engelder and Engelder, 1977; Slaughter, 1982; Geiser, 1988a, 1988b) (Fig. 5). Strain ellipticities for these samples (R_{yz}) range from 1.01 to 1.21 with a mean of 1.12 ± 0.04 and reveal diminishing magnitudes of LPS toward the foreland (Engelder and Engelder, 1977; Slaughter, 1982; Geiser, 1988a, 1988b) (Fig. 5). The long axes of the strain ellipses parallel the strike of folds.

The bulk finite strain in the remaining 23 nonfossiliferous samples (22 strained and 1 unstrained) throughout the Valley and Ridge Province is constrained using the normalized Fry method (Erslev, 1988) on detrital grains (Fig. 5). We found a good match between the mean ellipticity of bedding-plane strain (R_{yz}) between both the deformed crinoid ossicles and the quartz grains, as well only minor variation in the magnitude of XZ and YZ. Thus, for the purposes of our analysis, we limit our discussion of bulk finite strain to the mean ellipticity (R_{xz}) at each site. Mean ellipticity values range from 1.13 to 1.28 with a mean of 1.21 ± 0.04 (Table 2; Fig. 5). We observe no correlation among calculated R_{xz} values with respect to their position within the orogen, proximity to mapped structures, grain size, stratigraphic position, or orientation of the



A. Thickness of the sedimentary package above the basement and 3° slope of basement (top of Cambrian Waynesboro Formation) determined from industry seismic reflection lines (Gwinn, 1970; Beardsley and Cable, 1983; Scanlin and Engelder, 2003a).

B. Duplexing of Cambrian-Ordovician carbonates beneath the Valley and Ridge are inferred to fill space between the sole thrust and the roof thrust in the Ordovician Reedsville Formation. Duplexes are positioned such that the hinge zones of horses correspond with the positions of first-order anticlines on the surface, as per the fault-bend model (Suppe, 1983).

C. Duplexes 11 and 12 fill space beneath the southern most syncline in the Valley and Ridge where the fold-thrust belt is yielding to a different style of deformation dominated by extensive pressure solution and ≥50% LPS in the Ordovician Martinsburg Fm. (i.e., Wright and Platt, 1982).

D. Bedding thickness variations within the Ordovician Reedsville Fm are limited to the thickness necessary to satisfy map scale constraints. Thomas (2001, 2007) documents similar thickness variations from seismic reflection lines in the southern Appalachians of Alabama.

E. Detachment at the base of the Silurian Wills Creek Fm based upon field observations (i.e., Nickelsen, 1986; Klawon, 1994). The Wills Creek detachment feeds displacement from horses 8 and 10 onto the detachment in the Silurian Salina Group on the Appalachian Plateau (i.e., Prucha, 1968; Wiltchko and Chapple, 1977; Davis and Engelder, 1985).

F. Nittany Anticlinorium

G. Location and depth of thrust faults in fault-propagation anticlines are inferred to accommodate shortening in cores of both anticlines.

H. Bed thickness variations on the Appalachian Plateau vary in the plastic Salina evaporite group (i.e., Prucha, 1968; Wiltchko and Chapple, 1977).

I. Bed thickness variations in the predominately folded sequence are constrained by local stratigraphic studies and regional scale correlations (i.e., Ayrton, 1963; Colton, 1963; Hoskins et al., 1963; Sutton, 1963; Gwinn, 1964; Dyson, 1967; Faill and Wells, 1974; Hoskins, 1976; Faill et al., 1977, 1978; Berg and Edmunds, 1979; Wells and Bucek, 1980; Inners, 1981, 1997; Berg et al., 1983, 1993; Epstein, 1986; Edmunds, 1993, 1996.

J. Position of the New York State line in the restored section assuming 13% LPS along the length the of the cross section from the pin line at the northern end of the line of section (Figure 1).

Plate 1. (A) Simplified geologic cross section extending from the southern boundary of the Valley and Ridge physiographic province to the pin line beyond the extent of documented layer-parallel shortening on the Appalachian Plateau. Area of B is shown within the red box. (B) Balanced cross section from A–A′ extending from the southern boundary of the Valley and Ridge physiographic province to the Pennsylvania–New York border in the undeformed foreland. (C) Restored cross section. The deformed length (271 km) and restored undeformed length (348 km) are measured from the pin line beyond the extend documented layer-parallel shortening on the Appalachian Plateau, resulting in 77 km of shortening, or 22%. LPS—layer-parallel shortening. Circled numbers 1–12 are faults (see text and Fig. 11). To view the full-sized PDF file of Plate 1, please visit http://dx.doi.org/10.1130/GES00676.S1 or the full-text article on www.gsapubs.org.

TABLE 1. SHORTENING DATA FROM DEFORMED CRINOIDS IN THE DEVONIAN TRIMMERS ROCK FORMATION

Sample*	Latitude (°N)	Longitude (°W)	Ossicles measured	Bedding orientation†	Harmonic mean of R_s§	Mean ϕ trend (°)
1	40°19′20″	77°01′48″	37	45, 150	1.19 ± 0.04	073 ± 13
5	40°26′33″	77°06′32″	28	51, 161	1.23 ± 0.05	068 ± 15
7	40°29′10″	77°01′51″	48	81, 169	1.26 ± 0.04	070 ± 21
14	40°38′53″	76°57′21″	51	27, 129	1.23 ± 0.05	081 ± 19
20	40°47′13″	76°49′51″	44	28, 153	1.22 ± 0.07	063 ± 10

*Samples arranged as a function of position along the line of section from the hinterland to foreland.
†Dip, dip direction format for bedding orientations.
§R_s—ellipticity.

semimajor axis with respect to bedding (Table 2; Fig. 5; Supplemental File[1]). The outstanding nonfossiliferous sample, sample 26 (Table 2), was collected from an oolitic horizon at the base of the Ordovician Linden Hall Formation (Trenton Group) exposed in Nippenose Valley (Fig. 5). This sample is from below the Ordovician Reedsville detachment horizon and records a finite strain of 1.03, suggesting that LPS shortening is confined to the units above the Reedsville Formation. This measurement is consistent with 28 previously published measurements of grain-scale strain constrained by calcite twin data collected from the imbricated Cambrian–Ordovician sequence throughout the Pennsylvania salient (Ong et al., 2007).

We combine our 28 strain measurements with 60 previously published measurements (53 measurements from crinoid ossicles and 7 from calcite twin data; Engelder and Engelder, 1977; Engelder, 1979b; Slaughter, 1982; Geiser, 1988a, 1988b) for a total of 88 individual measurements of grain-scale strain across the study area from the Valley and Ridge Province through the Appalachian Plateau. Within the Valley and Ridge, our strain values are based on a compilation of 22 new LPS measurements in quartz grains, 5 new distorted crinoid ossicles (Tables 1 and 2), and 7 published strain estimates. From these data we calculate a mean strain of 1.21 ± 0.04, with no systematic spatial or lithologic variability. Thus, for a balanced cross section, 20% LPS must be restored through the Valley and Ridge. This value is in agreement with quantified estimates of strain from Nickelsen (1983) and Spiker and Gray (1997), but is significantly lower (with respect to the shortening budget) than the 28% required in the cross section of Herman (1984). On the Appalachian Plateau, we calculate 13% LPS based on strain estimates from 53 samples containing crinoid ossicles (Engelder and Engelder, 1977; Slaughter, 1982; Geiser, 1988a, 1988b).

The majority of the strain ellipses from samples in the predominately folded cover sequence have major axes that are oriented normal to bedding, or plunge steeply to the south when bedding is restored to horizontal (Fig. 7; Table 2). This suggests that the finite strain recorded in the quartz grains most likely developed during the early stage of progressive deformation in the central Appalachian foreland (Gray and Mitra, 1993). In instances where the major axis of the finite strain ellipse is oblique to bedding, the orientation of the axis may be used to constrain the sense of shear. The plunge southward of the majority of the major axes is consistent with the top-to-the-foreland (north) shear sense documented by Gray and Mitra (1993). However, in five of the samples we analyzed, the major axis of the strain ellipse plunges to the north when bedding is restored to horizontal (samples 8, 10, 15, 24, and 27 in Table 2). We can relate these northward plunges to the local structural setting (Supplemental File [see footnote 1]). Samples 8 and 10 were collected in the immediate vicinity of the mapped north-vergent regional-scale Buffalo Mountain thrust fault (Hoskins, 1976). Sample 15 was collected from the hanging wall of a hinterland-verging wedge fault. Samples 24 and 27 were collected along the northern limb of a first-order anticlinorium in the northern Valley and Ridge and are consistent with flexural-slip folding. Although these few samples are affected by local processes, the majority of the samples analyzed are consistent with a top-to-the-foreland sense of shear throughout the Valley and Ridge.

CROSS-SECTION CONSTRAINTS

The cross section we present here (Plate 1) shares several commonalties with earlier geologic cross sections across the Pennsylvania salient. Like Herman (1984), we invoke a passive roof duplex solution, and based on 1.03 finite stain that we measured in an oolitic horizon of the Late Ordovician Trenton Group, we assume that the imbricated layer in the region of our section was not shortened by LPS strain mechanisms during the Alleghanian orogeny. This is consistent with calcite twin data collected from the imbricated sequence of Paleozoic carbonates below the Reedsville detachment throughout the Valley and Ridge Province of Pennsylvania (Ong et al., 2007).

TABLE 2. MICROSCOPIC STRAIN SAMPLES LISTED BY LOCATION AND ASSOCIATED VALUES

Sample*	Latitude (°N)	Longitude (°W)	Fm†	Bedding orientation§	Orientation of semimajor axis	Restored semimajor axis	φ**	R_{xy}††
2	40°23′33″	77°00′50″	Dciv	39, 172	65, 028	67, 133	73	1.25
3	40°30′32″	76°43′07″	Dccf	29, 351	60, 187	82, 261	90	1.22
4	40°27′45″	76°57′26″	Dciv	22, 160	64, 216	48, 193	56	1.17
6	40°29′24″	76°57′15″	Dh	35, 210	70, 270	42, 233	47	1.17
8	40°31′01″	76°59′09″	Dcsc	59, 182	30, 182	29, 002	30	1.14
9	40°31′14″	76°59′12″	Mp	73, 348	10, 348	63, 168	63	1.21
10	40°31′28″	76°58′12″	Mp	72, 000	60, 240	30, 330	41	1.23
11	40°33′28″	76°59′52″	Mp	60, 160	35, 340	85, 160	85	1.23
12	40°33′39″	76°59′35″	Mmc	51, 128	40, 195	10, 174	46	1.19
13	40°34′01″	76°58′40″	Mp	52, 150	70, 330	58, 150	58	1.20
15	40°37′08″	76°57′16″	Dh	51, 166	30, 358	71, 019	81	1.27
16	40°39′00″	76°56′40″	Dh	43, 006	30, 188	73, 192	73	1.25
17	40°39′55″	76°53′34″	Dh	35, 345	50, 165	85, 165	84	1.18
18	40°41′53″	76°47′25″	Dtr	04, 345	77, 192	80, 203	82	1.17
19	40°42′00″	76°51′00″	Dtr	06, 262	55, 155	56, 164	85	1.28
21	40°53′16″	76°50′34″	Dciv	42, 164	63, 323	71, 194	74	1.13
22	40°56′33″	76°50′45″	St	13, 355	35, 193	47, 197	51	1.24
23	41°00′45″	76°65′07″	Swc	49, 357	34, 152	98, 069	86	1.26
24	41°01′58″	76°58′35″	Sc	39, 350	48, 129	64, 070	86	1.22
25	41°05′46″	77°07′23″	St	54, 000	40, 340	12, 164	19	1.26
26	41°09′19″	77°10′49″	Oh	13, 013				1.03
27	41°13′39″	76°57′06″	St	49, 358	40, 167	81, 092	89	1.14
28	41°20′45″	77°05′33″	Dsk	14, 346	42, 150	55, 145	58	1.26

*Samples arranged as a function of position along the line of section from the hinterland to foreland.
†Formation—see Plate 1 for abbreviations.
§Dip, dip direction format for bedding orientations.
**φ—angle of inclination of the long axis of the best-fit strain ellipse.
††R_{xy}—ellipticity. Strain constrained using normalized Fry method (Erslev, 1988) to measure bulk finite strain.

[1]Supplemental File. PDF file of table summarizing the field relationships of samples where grain-scale strain was measured and representative samples illustrating grain-scale strain in non-fossiliferous samples. If you are viewing the PDF of this paper or reading it offline, please visit http://dx.doi.org/10.1130/GES00676.S2 or the full-text article on www.gsapubs.org to view the supplemental file.

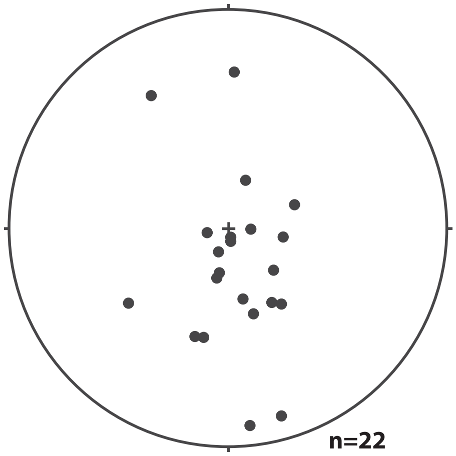


Figure 7. Equal-area lower-hemisphere stereonet for 22 measurements of bulk finite strain from the predominately folded cover sequence in the Valley and Ridge, showing orientations of the long axes for calculated strain ellipses after restoring bedding to horizontal.

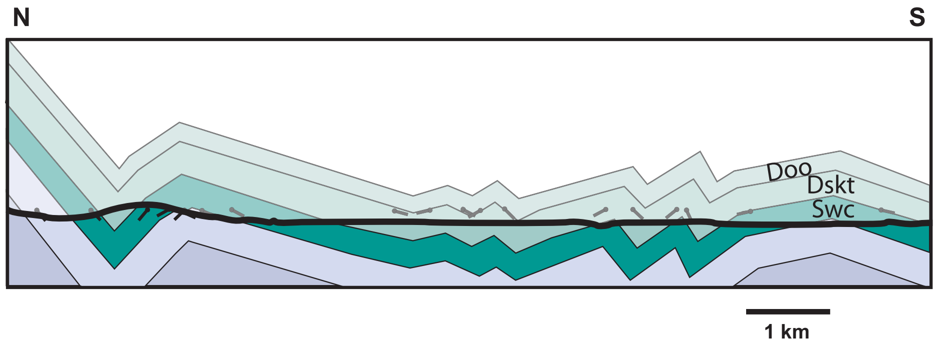


Figure 8. Detailed geologic cross section of second-order folds in the northern Valley and Ridge Province. The line of section is shown in Figure 4, and the stratigraphic color key is in Plate 1.

At the scale of the orogen, all shortening in this imbricated quartzite and carbonate layer is presumed to be accommodated solely by fracturing and faulting (i.e., Hatcher, 1989). A décollement horizon (or roof thrust) separates the thrust repeated Cambrian–Ordovician strata from the folded Ordovician through Mississippian strata.

The décollement is a gently (3°) southeastward-dipping planar surface that increases from a depth of 7200 m at the Appalachian front to ~10,600 m beneath the southern margin of the Valley and Ridge (Gwinn, 1970). This generalized geometry is consistent with the three published industry-acquired seismic lines in the central Valley and Ridge (Pohn and Coleman, 1991) (Fig. 5). One of these lines crosses the line of section, providing a direct constraint on the depth of the décollement.

The 271 km cross section (Plate 1) was drawn at a scale of 1:100,000; thus dip data depicting outcrop-scale folding, or larger scale map patterns, are lost. Second-order folds contribute significantly to shortening in the cover sequence (e.g., Markley and Wojtal, 1996; Hogan and Dunne, 2001). To quantify this contribution, we looked for mapped regions of second-order folds below the scale of the cross section. We identified one 11.5 km segment of second-order folds immediately off the line of section (red line in Fig. 4) and constructed a cross section at a scale of 1:50,000. This larger scale cross section increases our shortening estimate by 1.7 km (13%) (Fig. 8). Our shortening estimates do not include shortening accommodated by outcrop-scale structures such as folds and wedge faults (Fig. 9). To the south in the West Virginia portion of the Valley and Ridge, quantification of outcrop-scale structures contributes 10% (4.7 km) of shortening (Hogan and Dunne, 2001). Our examination of all roadcuts, stream cuts, and excavations along the line of section during winter months when vegetation was not obscuring exposures indicated that outcrop-scale structures are markedly less common in this section of Pennsylvania than farther to the south. However, since we cannot identify every mesostructure and not including these structures limits the calculated shortening, we emphasize that the shortening calculated is still a minimum.

BALANCING THE CROSS SECTION

We begin the process of balancing the cross section by restoring the folded bed length and accompanying LPS in the Appalachian Plateau to its original 185.2 km length (Plate 1). This restoration places 14 km of Appalachian Plateau strata south of the Nittany Anticlinorium (F in Plate 1), which marks the northern boundary of the Valley and Ridge and is the location where the décollement climbs from the Cambrian Waynesboro Formation to the Silurian Salina Formation (Wills Creek equivalent), or the salt décollement, of the Appalachian Plateau (Faill et al., 1977; Beardsley and Cable, 1983; Beardsley et al., 1999; Scanlin and Engelder, 2003b). We depict the Nittany Anticlinorium as a fault-bend fold with the northern-dipping limb as the hanging-wall ramp on the Silurian footwall flat, and the southern limb tilted by the corresponding footwall ramp for the fault-bend fold. The restoration of both folded and LPS Appalachian Plateau strata south of the Nittany footwall ramp requires that either (1) the northernmost footwall ramp that climbs from Cambrian to the Silurian strata is 15 km farther to the south, or (2) the Appalachian Plateau strata were shortened via LPS above the Silurian Salina Formation and before folding of the Valley and Ridge. Moving the footwall ramp to the south would require a double thickness of Cambrian to Ordovician strata between the folded cover sequence and the décollement, which is not possible with the available space. Thus, in order to place the northernmost footwall ramp under the Nittany Anticlinorium, and account for shortening on the Appalachian Plateau, we propose the following kinematic scenario to balance the shortening budget. Two of the southernmost horses (8 and 10, Plate 1) cut upsection from the Cambrian Waynesboro Formation to the Silurian Wills Creek Formation, making the Wills Creek Formation the décollement horizon for 14 km of slip transferred from the duplex out onto the Appalachian Plateau. This deformation predates the formation of horses 1–7 and the resulting Valley and Ridge (E in Plate 1). The structural elevation of the two anticlines overlying horses 8 and 10 combined with the depth of the basal décollement (which increases to the south) leaves only these structures with sufficient space to repeat the entire Cambrian–Silurian section. Support for the Wills Creek Formation being an important décollement horizon and facilitating translation of strata onto the Appalachian Plateau is found in detailed studies of the Silurian shale, which indicates consistent top-to-the-foreland shear that predates folding, as well as disharmonic folding across the proposed detachment horizon (Klawon, 1994). Similar bedding-parallel detachment surfaces are mapped in the same interval throughout the Pennsylvania salient, although the definitive, detailed kinematic investigations required to determine the sense of displacement have not been completed (Berg et al., 1980). Detailed mapping in Tennessee indicates that the ~30-km-wide Cumberland Plateau represents a thin sheet that was

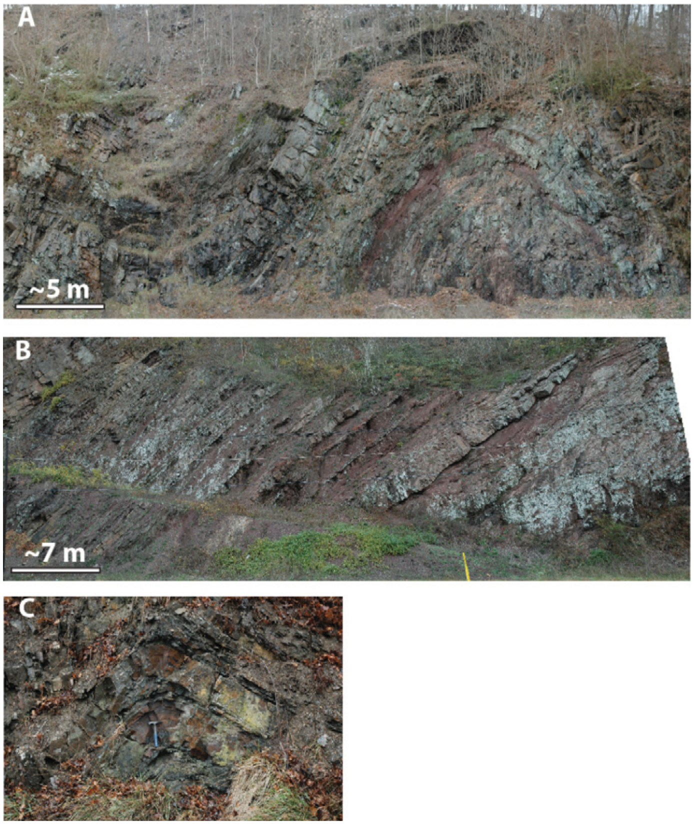


Figure 9. Outcrop-scale structural styles along roadcuts. (A) The most extensive example of third-order folds observed in the study area. Here, folds have wavelengths of tens of meters. (B) Homoclinally dipping package of Devonian Irish Valley Member exposed in a roadcut along the west shore of the Susquehanna River. Uniformal dipping sequences are typical throughout the study area. (C) Mesoscale anticline in the Devonian Tonoloway Formation. Note the well-developed wedge faults in the core of the fold in the competent beds and extensive axial planar cleavage in the less competent units. Note hammer for scale.

translated toward the foreland on a shallow subhorizontal detachment within Mississippian and Pennsylvanian strata (Milici, 1963); we propose a similar process for the Appalachian Plateau. The remaining shortening recorded on the Appalachian Plateau is from displacement on the final Valley and Ridge structure (horse 1; F in Plate 1). This fault translated 10 km of slip onto the Appalachian Plateau, creating the Nittany Anticlinorium, and 1 km of slip into the pair of tight anticlines at the Appalachian structural front (Plate 1). We infer small blind thrust faults (accommodating ~1 km of total slip) in the cores of both of these anticlines.

The region between the Nittany Anticlinorium in the north, and the early horses of the Valley and Ridge in the south (horses 8–10, Plate 1), is filled with horses of uniform thickness that repeat the Cambrian–Ordovician strata with the main sole thrust in the Cambrian Waynesboro Formation and the main roof thrusts in the Ordovician Reedsville Formation (Fig. 3 and Plate 1). These horses are positioned such that the hinge zones of the horses corresponded to the positions of large-scale anticline cores exposed at the surface, as per the fault-bend folding model (Suppe, 1983). Smaller scale anticlines with 1–3 km wavelengths do not match the geometry of the underlying horses, and are inferred to represent detachment folds above the roof thrust in the Reedsville Formation. The thickness of the Reedsville Formation, which includes the upper detachment for the imbricated carbonate layer, varies greatly along the line of section. Although measurements of thickness variations in the Reedsville Formation along the line of section are lacking, we assume that the variations are limited to those necessary to avoid space problems above the roof thrust. Support for large thickness variations in weak units throughout the Appalachians can be found in industry seismic data from the southern Appalachians fold-thrust belt to the Black Warrior foreland basin in Alabama that highlight locally thick zones of folded cover strata that are ductilely thickened along décollement horizons (Thomas, 2001, 2007).

Horses 11 and 12 are inferred in order to fill space beneath the southern limb of the southernmost syncline in the Valley and Ridge, as well as to balance the amount of shortening in the folded strata.

ESTIMATES OF SHORTENING

Balanced Cross-Section Estimates

Restoration of the balanced cross section provides the minimum estimate of horizontal shortening along the transect A–A′ through the Pennsylvania salient (Table 3). The shortening estimates represent a summation of deformation at the grain and map scale, but do not include outcrop-scale deformation. The deformed crinoid ossicles and quartz grains record predominately early-stage LPS. Many of the analyzed quartz ellipses are tilted to the north, indicating a top-to-the-north shear sense that postdates LPS but predates map-scale folding. This sequential development of microstructures to macrostructures is similar to that identified in other areas throughout the Pennsylvanian

TABLE 3. SHORTENING ESTIMATES ALONG THE CROSS SECTION A–A′

		Length		Shortening	
		Original (km)	Final (km)	(km)	(%)
Appalachian Plateau					
	Cover layer	185	161	24	13
	Stiff layer	161	161	0	0
Valley and Ridge					
	Cover layer*	163	110	53	32
	Stiff layer	187	110	77	41
Total					
	Cover layer	348	271	77	22
	Stiff layer	348	271	77	22

*Cover layer shortening in the Valley and Ridge is accommodated by layer-parallel shortening (33 km) and map-scale features (20 km).

salient (Nickelsen, 1979; Gray and Mitra, 1993; Faill and Nickelsen, 1999). The map-scale folds record the latest deformational event. Synchronous with the large-scale folding are local rotations of quartz grains due to flexural-slip faulting and wedge faulting during fold formation. The shortening magnitude across the Appalachian Plateau is almost entirely taken up by LPS. The 13% LPS (i.e., Engelder, 1979a; Geiser and Engelder, 1983) and restoring the broad gentle folds mapped by Wedel (1932) yields 24 km of shortening of the Appalachian Plateau strata. Shortening estimates across the Valley and Ridge also contain a significant component of LPS combined with map-scale folds. In this region, a total of 53 km of shortening is partitioned between 33 km of LPS and 20 km of shortening due to map-scale folding. Of the 20 km of map-scale folding, 2 km of that is from the larger scale, second-order folds (discussed in Cross Section Constraints). The total shortening for the cross section is 77 km (22%).

Area Balance Estimates

The area between the basal and upper detachment divided by the thickness of the strata that are duplicated within that area yield the original length of the deformed section and a shortening amount (Mitra and Namson, 1989). In Figure 10 we illustrate three area-balancing scenarios. The first scenario highlights the area of repeated Cambrian through Ordovician rocks from our balanced cross section (Plate 1) in black. The area of this cross section, divided by the thickness of the strata provides an undeformed length (l_o; Fig. 10). The difference between l_o and the deformed length (l_f) is the magnitude of shortening. Not surprisingly, this length of 76 km is essentially equal to the 77 km of total shortening calculated for the cross section (Table 3). Note the amount of area highlighted in Figure 10A is strongly dependent on the kinematic scenario we proposed where two faults (faults 8 and 10 in Plate 1) repeat the Cambrian–Silurian section. If we assume that the décollement in the Ordovician Reedsville Formation (Fig. 3 and Plate 1) always separates folded strata from faulted strata, and is never cut by thrust faults except at the Nittany Anticlinorium, then we can calculate two additional shortening estimates based on the excess area between the top of the Cambrian–Ordovician section and the base of the folded Reedsville Formation. The second scenario uses the same geometry just below the Wills Creek detachment as scenario 1, but assumes that the Reedsville detachment is not cut by thrust faults and all of the area beneath it is filled by repeating the Cambrian–Ordovician section. Shortening in scenario 2 (Fig. 10B) is 86 km because there is more space that must be filled by the repeating section. It also requires that at least some of the space beneath the structural high at the southern end of the section (highlighted by the extent of Ordovician Martinsburg Formation

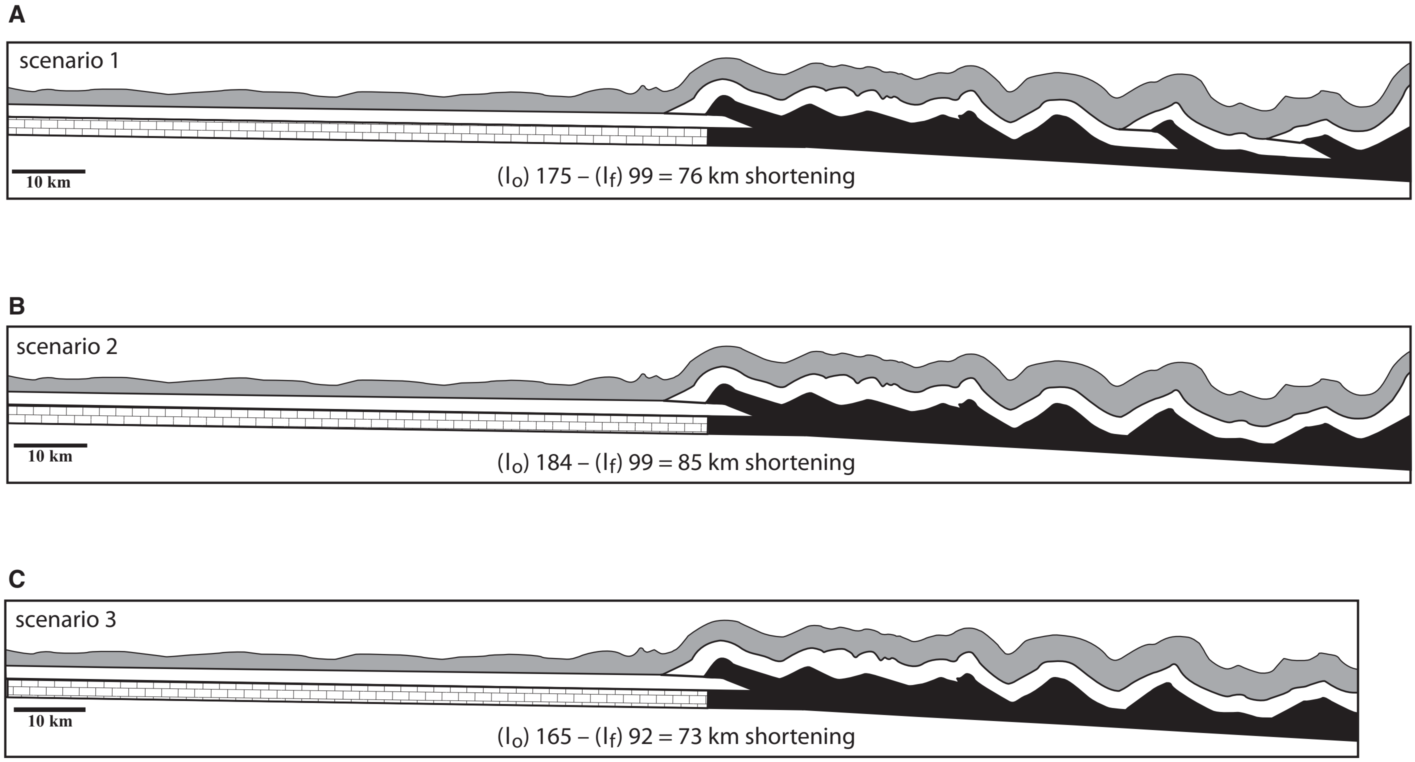


Figure 10. Area-balance calculations for the Cambrian–Ordovician sequence along profile A–A′. Symbols: black shaded area—deformed Cambrian–Ordovician strata; patterned region—undeformed Cambrian–Ordovician strata; white region—strata between the Ordovician Reedsville Formation and the Silurian Wills Creek Formation; gray shaded region—strata above the Silurian Wills Creek Formation; l_o—undeformed length; l_f—deformed length. (A) Area of repeated Cambrian–Ordovician rocks from our balanced cross section (Plate 1). Note the amount of area highlighted in A is strongly dependent on the kinematic scenario we proposed where two faults (faults 8 and 10 in Plate 1) repeat the Cambrian–Silurian section. (B) Area balance assuming that the décollement in the Ordovician Reedsville Formation (Fig. 3 and Plate 1) always separates folded strata from faulted strata and all space is filled by repeating the Cambrian–Ordovician section. (C) Area balance assuming that the southern limit of the cross section is situated along the axis of the southernmost syncline. Here, the cover rocks suggest 67 km of shortening, while the area balanced Cambrian–Ordovician section indicates 73 km of shortening.

at the surface) is filled with repeated Cambrian–Ordovician rocks, and as a result of the extra shortening in the Cambrian–Ordovician section, there is ~10 km of unaccounted shortening in Ordovician and younger rocks. This shortening could be due to failure to account for outcrop-scale structures, strain, or a combination of both. To place this in perspective, Hogan and Dunne (2001) calculated 4.7 km of outcrop-scale strain along a 48-km-long section of the Valley and Ridge Province in West Virginia. The 10 km of outcrop-scale strain along our 110-km-long section is comparable in magnitude. To account for the discrepancy via microstrain alone would require 25% strain through the Valley and Ridge, but only 8 of 28 samples analyzed show values as high as 25% ± 1% strain. The third scenario (Fig. 10C) does not require the structural high at the southern limit of our cross section to be filled with a repeated Cambrian–Ordovician section. This structural high could be filled with basement and Cambrian rocks that are structurally lower than the Cambrian Waynesboro décollement through the Valley and Ridge and exposed to the north and south in the Great Valley (Berg et al., 1980). Shortening in these rocks would equal that documented in the Valley and Ridge and not add to it. In this example, we place the southern limit of the cross section at the hinge of the southernmost syncline; here the cover rocks suggest 67 km of shortening while the area balance indicates 73 km of shortening. If we add the additional 10 km of cover rocks to account for outcrop-scale shortening, and 4 km to the faulted sequence to account for the truncated edge of horse 10, we can balance the section with 77 km of shortening by filling the structural high with faulted basement. Evaluating these scenarios rests on the sequential kinematic development of the central Appalachian fold-thrust belt and is discussed in the following.

SEQUENTIAL DEVELOPMENT OF THE PENNSYLVANIA SALIENT

Given that the structures drawn on a geologic cross section have kinematic significance, the kinematic admissibility of a balanced geologic cross section can be tested by sequentially deforming or retrodeforming the section (e.g., Geiser, 1988b; Evans, 1989; McQuarrie, 2002; Robinson, 2008). Forward modeling takes the retrodeformed cross section and attempts to produce the geometries depicted in the balanced section by successively moving successive thrust sheets as prescribed by the user. Alternatively the deformation depicted in the cross section can be sequentially removed to create a series of sequential deformation steps. We used 2DMove to sequentially undeform the cross section, but discuss the sequential development moving forward (in time).

The proposed sequential development of the central Appalachian fold-thrust belt (Fig. 11) is based on the balanced geologic cross section (Plate 1). The space between the orange cover sequence and the Pennsylvania–New York boundary line, as well as the gap in the cover sequence in Fig. 11A at the Appalachian front, equals the amount of shortening accommodated by LPS on the Appalachian Plateau.

The total restored fold-thrust belt from the pin line in New York State measures 348 km and the southernmost 277 km (from the New York State–Pennsylvania border) is shown in Figure 11A. The void spaces above the roof thrust in the Reedsville Formation (upper thick black line in Fig. 11) represent 20% LPS in the cover sequence (blue and brown lines in Fig. 11) in the Valley and Ridge and 13% accommodated within the cover sequence on the Appalachian Plateau. Between the time steps shown in Figures 11A and 11B, the cross section is shortened by 27 km (8%). In the imbricated carbonate sequence, this shortening is accomplished by in-sequence slip along the sole thrust of horse 12 (7.6 km), 4.9 km of slip along the sole thrust of horse 11, 7.4 km of slip along the sole thrust of horse 10, 1.0 km along the sole thrust of horse 9, and 6.4 km of slip on the sole thrust of horse 8. The overlying cover sequence is folded into forced fault-bend folds above the hanging-wall cutoffs in the imbricated sequence. Horses 10 and 8 shorten Cambrian–Silurian strata and translate ~14 km of Silurian and younger strata onto the Appalachian Plateau via the Wills Creek–Salina detachment (thin black line, Fig. 11B). In the interval between 11B and 11C, the total length of the cross section decreases by 21 km, from 321 to 300 km. This shortening is accomplished by in-sequence slip along the sole thrusts of horses 7 (6.4 km), 6 (4.1 km), 5 (1 km), and 4 (9.4 km), and folding of the cover sequence northward to the hanging-wall cutoff of horse 4.

The interval between Figures 11C and 11D represents an additional 29 km (10%) shortening resulting in the present geometry (Fig. 11D). This shortening is accommodated by in-sequence slip along the sole thrust of horse 3 (10.3 km), horse 2 (7.4 km) accompanied by detachment folding of the Ordovician and younger strata, and 9.5 km of slip along the sole thrust of horse 1, which feeds displacement onto the Appalachian Plateau and to each of the blind thrust faults in the cores of the two southernmost anticlines at the Valley and Ridge–Appalachian Plateau transition.

DISCUSSION

The kinematic scenario presented here requires the Silurian detachment to act as a weak décollement layer that transfers slip from older southern horses on to the Appalachian Plateau and concentrates the LPS only in the Appalachian Plateau region. The weak Salina salts can efficiently decouple deformation above and below this horizon, and transfer slip to the northern limit of the salt. Physical analogue models of salt décollements show that deformation structures do not sequentially propagate, but rather jump to the frontal salt pinch-out and then internally shorten (Costa and Vendeville, 2002). In addition, strain studies of detachment folds in the southern Pyrenees of Spain show that LPS deformation processes are only present above the salt detachment. Samples below the salt detachment are essentially unstrained (Sans et al., 2003). How LPS is partitioned through the orogen with time, following the same kinematic scenario as outlined herein, is illustrated in Figure 12A.

Alternatively, we can evaluate the kinematic scenario required by area balance 2 and 3 (Figs. 10B, 10C). In these scenarios, the faulted layer is confined to the Cambrian–Ordovician section and the Late Ordovician–Silurian rocks deform by folding and LPS. Here, the Ordovician Reedsville Formation efficiently decouples rocks that undergo 20% LPS from strata that are essentially unstrained (Table 2, sample 26). LPS precedes deformation via folding and migrates northward with time (Fig. 12B). LPS on the Appalachian Plateau is a continuation of this process with 14 km of LPS occurring in Ordovician–Mississippian strata. However, the last 10 km of LPS and broad folding of the Appalachian Plateau is due to the frontal-most thrust of the Valley and Ridge climbing upsection into the Silurian Salina Formation, and the 10 km of motion on this thrust must be balanced by folding and LPS above the Silurian salts.

Accurately discriminating between these two kinematic scenarios would require drill hole information beneath the salts on the Appalachian Plateau to evaluate whether the rocks are strained (scenario 2) or not (scenario 1), or very detailed seismic in the region of horses 8 and 10 (Plate 1). However, due to the fairly abrupt change in strain values between the Valley and Ridge and the Appalachian Plateau, versus a gradual decrease in values, we suggest that not only are the LPS values different, but the kinematics by which these rocks shorten are different. Thus, our preferred scenario is scenario 1, where early shortening in the Valley and Ridge is transferred out along a weak décollement in the Silurian Salina Formation.

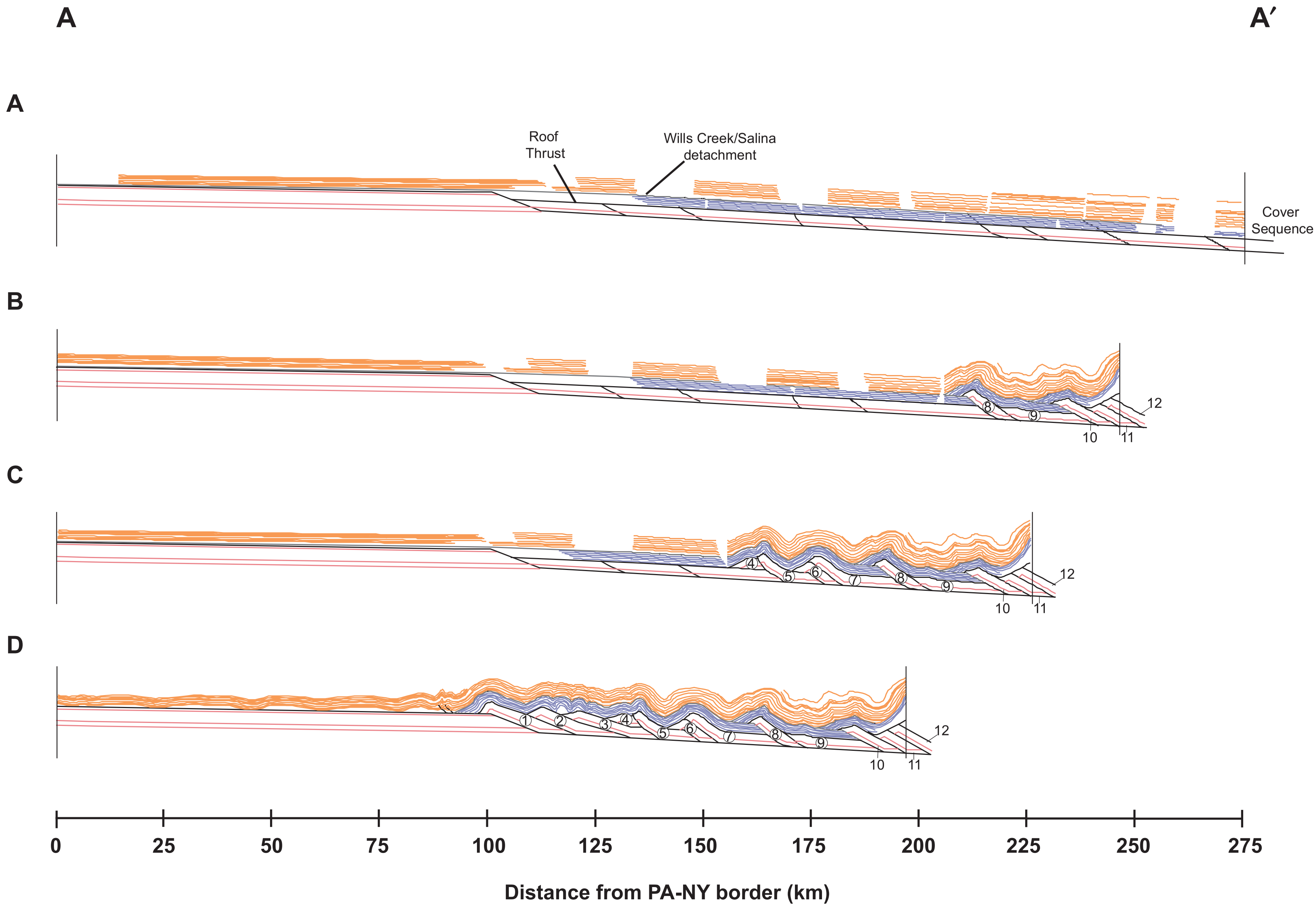


Figure 11. Kinematic evolution diagram showing the structural development of cross-section A–A′. The cross section (Plate 1) is sequentially restored in four steps. Void spaces within the cover sequence correspond to line length lost via layer-parallel shortening. Black—thrust faults in the imbricated carbonate sequence (labeled 1–12; see text); gray—Wills Creek–Salina detachment; pink—Cambrian-Ordovician contact in the mainly thrust-faulted sequence; blue—cover sequence strata (older than the Silurian Wills Creek Formation); brown—cover sequence strata (Wills Creek and younger). Scale is in kilometers with no vertical exaggeration. NY, PA—New York, Pennsylvania.

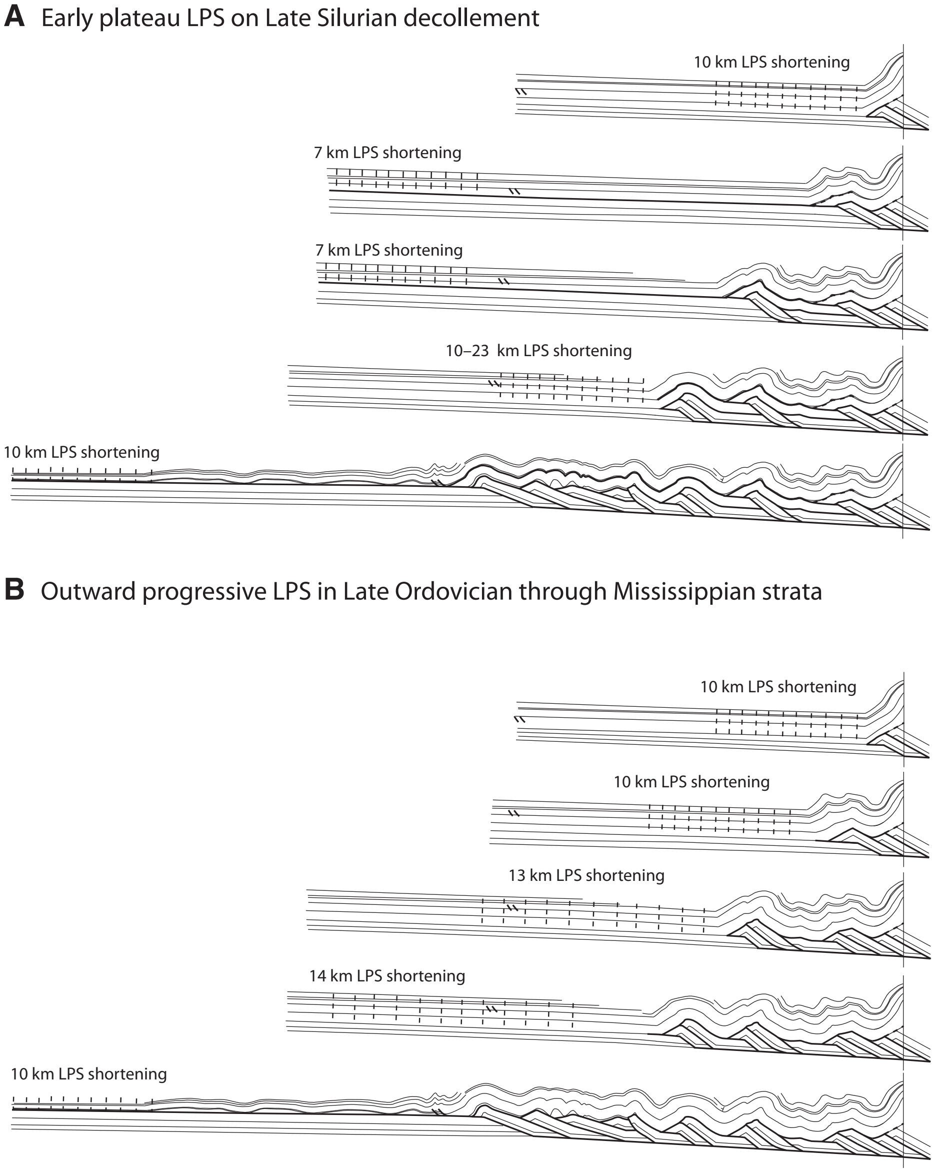


Figure 12. (A) Layer-parallel shortening (LPS) on the Appalachian Plateau is limited to strata above the Silurian Wills Creek Formation. This shortening is linked to early duplex development where two faults cut through the entire Cambrian through Silurian section. (B) LPS immediately precedes folding and only the last LPS (10 km) on the Appalachian Plateau is limited to Silurian and younger strata. In this scenario, the Ordovician Reedsville Formation always separates faulted strata from folded strata, and some LPS (14 km) is accommodated in the Ordovician and younger rocks in the plateau.

CONCLUSIONS

Our analysis emphasizes that magnitude and location of strain is a function of the kinematics of the fold-thrust belt. Thus, the addition of mesoscale to microscale structures into shortening estimates cannot simply be an add-on, but should accompany a kinematic model that illustrates the spatial and relative temporal distribution of shortening. By applying this to a balanced cross section through the Pennsylvania salient, from the southern margin of the Valley and Ridge to the northern limit of documented LPS on the Appalachian Plateau, we show that a sequential restoration of the cross section, combined with measurements of LPS, highlights how LPS is distributed above and ahead of thrust and fold shortening. In the Valley and Ridge, 77 km (41%) shortening within the mainly thrust-faulted Cambrian–Ordovician sequence is accommodated by development of a duplex. The overlying cover sequence accommodates 53 km (32%) of shortening by LPS and folding. The thrust-faulted carbonate sequence is not shortened across the Appalachian Plateau, where there is 24 km (13%) of shortening in the cover sequence. For the Appalachian Plateau to accommodate shortening and help alleviate the discrepancy between shortening of the mainly folded cover and imbricated carbonate sequences, 14 km of slip must be fed along an early detachment in the Wills Creek Formation in the Valley and Ridge onto a subhorizontal detachment in the weak salts of the Salina Group on the Appalachian Plateau. The Appalachian fold-thrust belt largely developed in sequence with slip from each consecutive horse in a growing duplex feeding slip into folding and LPS in the overlying cover rocks. This study shows that the amount of shortening needed to fill space between a seismically imaged detachment and the mapped cover sequence can be balanced by integrating regional macroscale and microscale structures and documented magnitudes of LPS.

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