Complex Proterozoic Crustal Assembly of Southwestern North America in an Arcuate Subduction System: The Black Canyon of the Gunnison, Southwestern Colorado

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The dominant orogenic fabric in Proterozoic rocks of the southwestern U.S. includes a series of NE-striking shear zones that are commonly interpreted as suture zones across which blocks of juvenile crust were assembled to the southern margin of Laurentia. New structural and geochronological data from southwestern Colorado suggest that fabrics related to assembly of tectonostratigraphic terranes in this area strike northwest. The NW-striking foliations represent deformation at ca. 10–20 km paleodepths (ca. 1.77–1.71 Ga), and are parallel to magnetic anomalies and to gradients in mantle velocity structure. The agreement between these data sets suggests that the NW-striking structures are important at lithospheric scale, extend to >100 km depth, and may record assembly of southwestern Colorado across NW-striking tectonic boundaries. Geochronologic data indicate that northwest (central Colorado)-and northeast (Cheyenne belt)-striking boundaries developed simultaneously during accretion of southwestern Laurentia between ca. 1.78–1.73 Ga. We propose that the Yavapai province at ca. 1.75 Ga may have involved a complex arcuate subduction system, with multiple arcs, analogous to that of the modern Banda Sea, in the Indonesia region.

INTRODUCTION

Models for forming juvenile continental crust commonly involve the complex collisions of oceanic elements and their amalgamation to the margin of an existing continent [Hamilton, 1979; Snyder et al., 1996]. In the southwestern U.S., most models involve crustal growth from the northwest to southeast based on the dominant NE-striking orogenic fabric [Reed, 1987; Karlstrom and Bowring, 1988]. However, the Paleoproterozoic rocks exposed in southwestern Colorado add an important new element to these models; here the dominant...
fabric is NW-striking over a large region and geochronological data suggest that it formed at the same time that NE-striking shear zones were active elsewhere in the region. An accurate model for assembly of the crust in the southwestern U.S. must account for coeval NW-and NE-striking zones [Albin and Karlstrom, 1991]. The goals of this paper are to describe, and provide time constraints on the NW-and NE-striking fabrics in southern Colorado. We highlight NW-striking aero-

Figure 1. Proterozoic rocks exposed in Colorado. Boxes show locations of Plate 1 and Figure 2. Black bars represent the NW-trending F2 folds (~1.70–1.72 Ga) and gray bars depict NE-trending F3 folds. Granitic batholiths after Reed et al. [1987]. Shear zone locations after Tweto [1963].
magnetic and mantle tomographic trends that coincide with the NW-trending surface foliations, and thereby delimit a ~200 km long and >100 km deep domain that may represent a major lithospheric scale structural domain in southwestern Colorado.

This study seeks to unravel the ~350 m.y. long history and complex kinematics recorded by Proterozoic rocks of southwestern Colorado, specifically those exposed in the Black Canyon of the Gunnison [Hansen and Peterman, 1968]. New structural and geochronological work in the Black Canyon of the Gunnison presented here also leads to a better understanding of several aspects of the ca. 1.79 Ga-1.43 Ga history of the southwestern U.S., including ca. 1.4 Ga intracontinental tectonism related to A-type plutonism.

BACKGROUND

Proterozoic rocks in Colorado are part of the 1300-km-wide belt of juvenile crust that extends from the Cheyenne belt, Wyoming, to Sonora, Mexico (Figure 1). The area of the Cheyenne belt initially formed as a rifted margin (ca. 2.1 Ga) and then became the fundamental Proterozoic/Archean suture [Karlstrom and Houston, 1984; Duebendorfer and Houston, 1987; Tyson et al., 2002]. The Proterozoic juvenile terranes south of the Cheyenne belt contain metasedimentary and metavolcanic rocks intruded by several generations of intrusive rocks [Reed et al., 1987]. These rocks are believed to have formed in a complex series of arcs and oceanic terranes built on and tectonically juxtaposed with locally older (ca. 2.0–1.8 Ga) continental fragments [Reed et al., 1987; Karlstrom and Bowring, 1988; Bowring and Karlstrom, 1990; Ilg et al., 1996; Shaw and Karlstrom, 1999; Selverstone et al., 2000; Hill and Bickford, 2001]. Uncertainty remains regarding the paleogeography and original geometry of tectonic blocks, and the location of possible sutures south of the Cheyenne belt.

We use the following terminology for provinces and events. The term “Yavapai orogeny” is used for the long and complex series of collisions involving oceanic tectonic elements (arcs, arc basins, oceanic terranes) and their assembly to North America (ca. 1.8–1.7 Ga). The resulting “Yavapai province” includes pre-1.70 Ga predominantly juvenile crust that extends from the Cheyenne belt south to central New Mexico and central Arizona and west to the Mojave province [Karlstrom and Bowring, 1988; Karlstrom and Humphreys, 1998]. This series of events was followed by crustal stabilization that resulted in a regional unconformity and deposition of mature quartzites (Plate 1) [Williams, 1987; Soegaard and Eriksson, 1989]. Scattered remnants of these quartzites extend well northward (into Colorado) from the sites of thickest preservation in northern New Mexico (Figure 1) [Williams, 1987]. The term “Mazatzal orogeny”, refers to the deformation at ca. 1.65 Ga that resulted in the assembly of 1.68–1.65 Ga tectonic blocks to the southeastern margin of the Yavapai province along a complex, northeast-striking boundary [Shaw and Karlstrom, 1999]. After the Mazatzal orogeny at ca. 1.65 Ga there followed a ~200 m.y. period of cratonic stability and residence in the middle crust of the Proterozoic rocks that are presently exposed in the southwestern U.S. [Bowring and Karlstrom, 1990; Williams and Karlstrom, 1996]. This stability was disrupted by emplacement of ca. 1.45–1.35 Ga A-type granites and temporally associated regional transpressive deformation [Nyman et al., 1994].

Southwestern Colorado provides excellent exposure of Paleooproterozoic rocks that range in age from ~1.8 to 1.68 Ga and record a sequence of events related to the growth, stabilization, and reactivation of juvenile lithosphere (Plate 1). The Dubois (ca. 1.77–1.76 Ga) and Cochetopa (ca. 1.74–1.73 Ga) successions are sequences of calc-alkaline plutons (1.75 Ga and 1.72–1.70 Ga) and contain inherited zircons with ages of ~2.5 Ga and 1.87–1.84 Ga (Plate 1) [Bickford and Boardman, 1984; Knoper and Condie, 1988; Bickford et al., 1989; Hill and Bickford, 2001]. Nd models suggest that these rocks represent juvenile material often interpreted as arc components [DePaola, 1981]. The presence of inherited zircons supports models for the presence of older crustal material of ~2.0–1.8 Ga in the subsurface (Trans-Hudson/Penokean age) [Hill and Bickford, 2001; Hawkins et al., 1996]. The Black Canyon area hosts quartz-rich metasedimentary rocks that are in contact with the Dubois and Cochetopa metavolcanic successions (Plate 1) [Hansen, 1981]. The nature of contacts between these three lithostratigraphic successions was obscured by intense polyphase deformation and may involve depositional contacts and/or tectonic juxtaposition of “blocks” [Afifi, 1981; Shonk, 1984]. Proterozoic rocks are also exposed in the Uncompahgre Plateau; these consist of migmatic gneiss (ca. 1.74 Ga) intruded by a synkinematic monzogranite (1.72 Ga; Figure 2) [Livaccari et al., 2001].

Regional ~1.4 Ga intracratonic tectonism is recorded by the emplacement of the Curecanti and Vernal Mesa monzogranites, several generations of pegmatite dikes (Plate 1) [Hansen, 1964; Hansen and Peterman, 1968; Bickford and Cudzilo, 1975; Hansen, 1981], and significant deformation and metamorphism [Jessup et al., 2002a,b; Jessup, 2003]. The Curecanti monzogranite (1420 ± 15 Ma) is undeformed and was emplaced as sheets that crosscut the vertically foliated migmatic gneiss host rock [Hansen, 1964; Hansen and Peterman, 1968]. The Vernal Mesa monzogranite (1434 ± 2 Ma; dated and discussed below) of the Black Canyon is interpreted as syn-tectonic [Hansen and Peterman, 1968; Hansen, 1981], but the Vernal Mesa of the Uncompahgre
Plate 1. Generalized geologic map of the Black Canyon, Dubois and Cochetopa successions [Hansen, 1971, 1972; Hedlund and Olson, 1973, 1974a,b; Olson, 1975, 1976; Olson and Steven, 1976 a,b; Lafrance and John, 2001]. Geochronology after Bickford et al. [1989]. The Red Rocks fault is reconstructed for the structural analysis and foliation trajectory map. Inset A shows the position of the Proterozoic rocks on either side of the Red Rocks fault as they currently exist. Insets B, C, D show cartoons of key overprinting relationships.
The plateau (1430 ± 22 Ga) is interpreted as post-tectonic [Livacari et al., 2001]. The discrepancy between these two interpretations may reflect different levels of emplacement and/or partitioned deformation.
GEOMETRIC AND KINEMATIC ANALYSIS OF THE BLACK CANYON REGION

In order to elucidate the nature of regional tectonism as expressed in southwestern Colorado, the following section describes details of a geometric and kinematic analysis of the well-exposed Proterozoic rocks in the Black Canyon of the Gunnison. This section builds on excellent mapping by Hansen [1971; 1972] and distinguishes four phases of deformation. We differentiate these phases with an understanding that they may represent distinct events, a progression of deformation, or some combination of both. The prevalent structural character throughout the Black Canyon was imposed during the first (D1) and second (D2) phases of deformation, while the third (D3) and fourth (D4) phases of deformation are limited to the Black Canyon shear zone (Plate 1). The four phases of deformation are generalized as follows. The first phase of deformation (D1) is recorded by mesoscopic and macroscopic isoclinal F1 folds of bedding and associated S1 axial planar foliation and is preserved in the hinges of F2 folds where subsequent overprinting is weak (e.g., Crystal Creek synform, Morrow Point and Wildcat Gulch antiforms; Plate 1, inset C). The second phase of deformation (D2) is recorded by variably plunging macroscopic F2 folds and L2 stretching lineations that resulted in the regionally dominant S1/S2 composite foliation that strikes north to northwest (e.g. Plate 1; Crystal Creek synform, Morrow Point antiform, Curecanti antiform, Wildcat Gulch antiform, and Gold Basin antiform). D3 is recorded by NE-trending, upright F3 folds with sub-vertical NE-striking S3 axial planar foliation and subvertical L3 stretching lineations that are concentrated in a ~5-km-wide-high-strain zone (Plate 1). We term this high-strain zone the Black Canyon shear zone and propose that it may be the southern extension of the Colorado Mineral Belt shear zone system [Tweto, 1963; McCoy et al. 2001]. D4 is recorded by reoriented L3-lineations, rotated into parallelism with a gently SW-plunging L4 stretching lineation. This lineation is perpendicular to ~1.4 Ga pegmatite dikes and is interpreted to record a reactivation of the Black Canyon shear zone during dextral transpressional shear. In the following section we present detailed observations from four structural domains that provide the basis for the four phases of deformation proposed above.

The Crystal Creek F2-Synform (Domain 1)

Several folds that contain sub-horizontal, N-S trending fold axes are located in the northwestern part of the Black Canyon of the Gunnison, including the Crystal Creek synform (Plate 1, Domain 1, Fig 3a). The gneissic S1 foliation in the Pitts Meadow granodiorite and migmatized quartzofeldspathic paragneiss is folded by F2 folds and is therefore oriented N-S, subvertical, and subparallel to the axial plane of the F2

Figure 3. a) Equal-area projection of structural data from domain 1 (e.g. Crystal Creek synform). The poles to the S1 foliation (squares) define the great circle used to calculate a subhorizontal F2 fold axis (large gray circle) of 01°→004° (N=135). The L2 stretching lineations (small gray circles) have an average orientation of 12°→182° (N=38). The orientation of fold axes and lineations suggests that the extension direction, as recorded by stretching lineations, is subparallel to the fold axes. b) Equal-area projection of poles to the S3 foliation within domain 2 (e.g. the Black Canyon shear zone). The tight clusters of poles to S3 demonstrates a consistent shear plane orientation of 246°, 84° (N=40) as shown by the dashed great circle. c) Equal-area projection of lineations and fold axes within the Black Canyon shear zone. Solid squares represent the orientation of stretching lineations in the migmatitic gneiss near the contact with the Vernal Mesa monzogranite and define a cluster with an average orientation of 32°→243° (N=13). Open circles depict the fold axes orientation in the migmatitic gneiss that varies between 29°→246° and 13°→056° with a maximum towards the southwest (N=21). Lineations in the layered quartzitic gneiss are shown by the distribution of open crosses that vary between 33°→250° to 45°→050°. We interpret the variation in lineation orientation as a record of the rotation of linear elements towards vertical in the center of the shear zone. Here, the average orientation of lineations in the layered quartzitic gneiss is 84°→302° (N=42). Fold axes within the layered quartzitic gneiss are shown as black crosses and have an average orientation of 82°→055° (N=10). The subparallel orientation of fold axes and lineations provide evidence for the rotation of fold axes into subvertical near the center of the shear zone. d) Equal area projection of the average orientation of NW-striking pegmatite dikes (292°, 51°N), shear plane (224°, 74°N), and L4 stretching lineation (32°→243°). The intersection of the pegmatite dikes with the shear plane is ~90° from the movement direction (X-axis) as defined by stretching lineations. If the pegmatite dikes represent tensile openings perpendicular to the X-axis and are used to define the finite strain axes, they are consistent with dextral shear with a NW-side-up component. e) Equal-area projection of poles to S1 foliation measurements from domain 3 defines the great circle used to calculate a F2 fold axis (gray circle) for the Morrow Point antiform with an orientation of 27°→175° (N=75). f) Equal-area projection of poles to S1 foliation measurements from domain 4 that statistically define an F2 fold axis orientation (gray circle) of 44°→292° (N=27) for the Curecanti Needle antiform. g) Equal-area projection of poles to S1 foliation measurements that defines an F2 fold axis orientation (gray circle) of 50°→318° (N=189) for Gold Basin antiform.
folds (Plate 1) [Hansen and Peterman, 1968; Hansen, 1971, 1972]. The Pitts Meadow granodiorite (ca. 1.713 Ga) [Jessup et al., 2002b] was emplaced before or during amphibolite facies metamorphism and migmatization, as shown by diffuse contacts where metasedimentary gneiss grades subtly into gneissic granite [Hansen and Peterman, 1968]. On the western limb of the Crystal Creek (F2) synform the Pitts Meadow granodiorite is sheared sinistrally and has a penetrative L/S-tectonite (17°→201°) fabric that is axial planar to N-S-trending F2 folds. The S1/S2 composite foliation is defined by aligned mica and flattened quartz and feldspar; elongate quartz and feldspar rods define L2 stretching lineations that are sub-parallel to F2 fold axes (Plate 1). An early generation of pegmatite dikes (ca. 1.709 Ga) [Jessup et al., 2002b] is isoclinally folded and locally boudinaged along the S1/S2 composite fabric and suggests ~E-W shortening and ~N-S extension in the X-axis (17°→201°) (Figure 4a; see CDROM in back cover sleeve). Asymmetric pegmatite boudins along the S1/S2 composite foliation record sinistral strike slip that we interpret to have formed either pre-or syn-F2 folding (Figure 4b; see CDROM in back cover sleeve).

The Green Mountain antiform (F2) is southeast of the Crystal Creek synform (Plate 1). This area contains schist and migmatized gneiss, migmatized gneiss, and quartz-biotite-plagioclase-zogranite, two generations of pegmatite, layered quartzitic gneiss, and characterized by blocks and screens of gneiss within quartzofeldspathic gneiss that is several hundred meters wide (Plate 1) and is isoclinally folded and locally boudinaged along the S1/S2 composite fabric and suggests ~E-W shortening and ~N-S extension in the X-axis (17°→201°) (Plate 1). The axial surface of the Green Mountain antiform bends southwest as it approaches the Black Canyon shear zone (Plate 1).

The Black Canyon Shear Zone (Domain 2)

The ~4-km-wide, NE-striking, Black Canyon shear zone (Domain 2; Figure 3b) is a zone of intense NE-striking, subvertical foliation (246°, 86°NW) and migmatization. It deforms a variety of rocks including the Vernal Mesa monzogranite, two generations of pegmatite, layered quartzofeldspathic gneiss with an average fold axis of 17°→199° and average stretching lineations of 34°→233° (Plate 1). The axial surface of the Green Mountain antiform bends southwest as it approaches the Black Canyon shear zone (Plate 1).

The Vernal Mesa monzogranite (1434 ± 2 Ma; dated and discussed below) is a NE-trending, 1.5 x 6 km body on the northwestern margin of the Black Canyon shear zone. It has a strong NE-striking, dominantly magmatic foliation defined by aligned feldspar megacrysts. Solid-state fabrics in the pluton are present but uncommon, and indicate that some regional deformation outlasted pluton crystallization. The contact between the monzogranite and migmatized quartzofeldspathic gneiss is several hundred meters wide and characterized by blocks and screens of gneiss within granite that decrease in abundance away from the contact. Fold axes and lineations are parallel in the migmatitic gneiss near the pluton and define a maximum of 32°→243° (Figure 3c; Figure 5; see CDROM in back cover sleeve). NW-striking ~5-m-wide pegmatite dikes (292°, 51°N) crosscut both the Vernal Mesa monzogranite and migmatitic gneiss to the northwest and therefore were intruded during or after emplacement of the Vernal Mesa monzogranite. Equal-area projection of the shear plane, stretching lineations, and average orientation of pegmatite dikes demonstrates the geometric relationship between these features (Figure 3d). The well-developed stretching lineations form a maximum that is ~90° from the intersection of the pegmatite dikes and shear plane (Figure 3d).
Morrow Point Antiform (Domain 3)

The Morrow Point antiform (F2) is one of a series of open synforms and antiforms; it has a fold axis orientation of 27° to 175° (Figure 3e). The Morrow Point antiform (F2) is cored by layered quartzitic gneiss and overlain by a thick sequence of migmatitic sillimanite-bearing schist. On the eastern limb of the broad F2 antiform there is a nappe-style isoclinal F1 fold with a pervasive axial planar S1 foliation. The S1 foliation is recorded in the hinges of parasitic F1 folds where aligned muscovite is at a high angle to bedding. Pegmatite sills and leucosomes are localized along the ~N-S-striking S1 foliation and are folded by F2 folds indicating that a migmatization event was syn-or-pre-F2. This is important in documenting that migmatization of country rock took place during D1 (Pitts Meadow), D2 (Morrow Point), and D4 (Vernal Mesa).

Curecanti and Wildcat Gulch Antiforms (Domain 4)

The Curecanti antiform (F2) folds migmatitic gneiss with a pervasive S1 foliation and trends northwest with a fold axis of 44° to 292° (Figure 3f). The cuspatc, open hinge, steep-limbed style of F2 folds appears to be consistent between the Morrow Point antiform and these areas to the southeast. However the trend of the fold axes undergoes an important transition from N-S to NW-SE as first noted by Hunter [1925]. Continuing to the southeast towards the Dubois and Cochetopa successions, the next major structure that we mapped is the Wildcat Gulch antiform (F2; 13° to 111°) just northwest of the contact between the Dubois and Black Canyon successions (Plate 1). The Wildcat Gulch antiform (F2) has a broad hinge and steep limbs and folds layered quartzitic gneiss that contains a pervasive S1 foliation. Overprinting relationships between the two phases of deformation are preserved in the hinge region just north of Wildcat Gulch. Here the S1 foliation is parallel to the axial surface of isoclinal F1 folds that fold graded bedding (Plate 1, inset C).

THE DUBOIS AND COCHETOPA SUCCESSIONS

Based on the following synthesis of data we propose three generalized phases of deformation recorded by the Dubois and Cochetopa successions (Plate 1). The first phase (D1) includes F1 folds of primary volcanogenic features and is associated with a strong axial planar S1 foliation that is defined by the alignment of various minerals and dominates much of the region (Plate 1, inset D; e.g., Iris syncline, Beaver Creek anticline, Vulcan syncline, Gunnison annular complex) [Afifi, 1981; Shonk, 1984; Lafrance and John, 2001]. The second phase of deformation (D2) produced shallow to moderately plunging NW-trending F2 folds with steep limbs (e.g. Gold Basin antiform Figure 3g), Gunnison annular complex) [Afifi, 1981; Shonk, 1984; Lafrance and John, 2001]. F2 folds (and hence development of the regional sub-vertical NW fabric) formed between 1.70 and 1.713 Ga, based on the syn-kinematic granite of South Beaver Creek (1.70 Ga) and late/post-kinematic tonalite of Gold Basin (1.713 Ga) [Afifi, 1981; Shonk, 1984; Bickford et al., 1989; Wortman and Bickford, 1990; Lafrance and John, 2001; Hill and Bickford, 2001]. The syn-to-post-kinematic tonalite of Gold Basin is largely undeformed but locally preserves a deformational fabric that is recorded by boudinage of some small dikes and attributed to the assembly of another arc [Wortman and Bickford, 1990]. Hinge zones of F2 folds are not as common as the steep limbs (cuspate-style), and therefore the pervasive fabric of the region is considered to be a subvertical S1/S2 composite fabric (Plate 1). The composite S1/S2 fabric was overprinted during a third event (D3), to create complex, more localized F3 folds and S3 foliations such as the southern limb of the Iris syncline (Plate 1, inset D) [Afifi, 1981]. At this location, the isoclinal Iris syncline folds primary features. During D2 this structure was refolded by a NW-trending F2 fold. On the southern end of the Iris synform a third phase of deformation is recorded by a NE-trending fold with a subvertical axial planar foliation (Plate 1, inset D) [Afifi, 1981].

THE BLACK CANYON AND DUBOIS SUCCESSION CONTACT

The contact between the rocks of the Black Canyon and Dubois succession provides a small field area to test various hypotheses for the assembly of the southwestern U.S. during the Paleoproterozoic. Here, the Proterozoic rocks of the Black Canyon share a pervasively sheared contact with the Dubois succession (Plate 1). We mapped the shear zone to determine whether the contact represents the juxtaposition of unrelated “blocks”, a sedimentary contact, or both.

The area is divisible into three main-tectonic units; the Proterozoic rocks of the Black Canyon (originally named the Black Canyon Schist), the Dubois succession (originally the Dubois Greenstone), and a lens of quartzite conglomerate that we here name the Cebolla Creek quartzite conglomerate [Olson and Hedlund, 1973]. As exposed near the contact, the rocks of the Black Canyon area are predominantly composed of various types of quartz-biotite schist/gneiss, quartzofeldspathic gneiss, amphibolite schist, and limited carbonate lenses, with little or no volcanogenic sediment. In contrast, the Dubois succession (1.77–1.76 Ga) has many features that are hallmarks of a volcanogenic origin [Afifi, 1981; Bickford et al., 1989]. These features include, but are not limited to, felsic and mafic metavolcanic rocks, pillow structures, amygdules, metachert, and flow rocks [Olson and Hedlund, 1973]. The
third unit is the Cebolla Creek quartzite conglomerate, as mapped by Olson and Hedlund [1973], which is folded with the Dubois succession and composed of rounded cobbles and pebbles of white to gray quartz with a quartz matrix (Plate 1).

The nature of the contact between the rocks of the Black Canyon and Dubois successions is not abrupt as originally mapped by Olson and Hedlund [1973], but a complex gradual transition marked by a progressive decrease in abundance of Dubois-succession-related rocks from the center of the field area to the north. There is a layer of marble near the northern limit of felsic volcanogenic rocks of the Dubois succession, an interesting and rare occurrence in the area. Further to the north, rocks of Black Canyon area, including many layers of amphibolite, dominate the area. We did not find any obvious primary features that demonstrate the protolith of these amphibolite layers.

The general structural fabric that dominates the area is a pervasive S1-foliation (280°, 67°N) and well-developed stretching lineation (60° → 310°). The S1 foliation and shear plane is also the axial planar foliation of isoclinal F1 folds and is refolded by open F2 folds of variable orientation. Many discrete mylonite zones along the S1 foliation record top-to-the-south/dextral oblique shear by C-C’ fabrics defined by feldspar porphyroclasts. These mylonite zones do not seem to crosscut the Cebolla Creek quartzite conglomerate.

Based on these criteria we propose that the original contact was likely sedimentary - not the contact between two juxtaposed “blocks” - and the pervasively sheared contact was the result of several phases of deformation. The timing of this movement is roughly bracketed between the deposition of the youngest metarhyolite (1.72 Ga) and the Cebolla Creek conglomerate.

THE CEBOLLA CREEK QUARTZITE CONGLOMERATE

The Cebolla Creek quartzite conglomerate is a synclinal “keel” of clean quartzite and quartzite conglomerate with well-preserved cross bedding that unconformably overlies the Dubois succession (Plate 1) [Olson and Hedlund, 1973]. Differences in foliation intensity and strain magnitude in the quartzite indicate that the quartzite was deposited upon the previously deformed Paleoproterozoic basement. The presence of an angular unconformity at the base of the quartzites agrees with observations from another quartzite ~10 km south [Hill and Bickford, 2001] and the nearby Needle Mountains where basement was deformed before and after quartzite deposition [Gibson and Harris, 1992].

We interpret the deposition of this quartzite as evidence for a syn-tectonic basin that developed after the assembly and unroofing of Yavapai province crust. The angular unconformity was subsequently buried and folded into a NE-striking syncline possibly due to shortening related to the Mazatzal orogeny. This quartzite is thus correlated with the Uncompahgre Group, Ortega quartzite, and Blue Ridge quartzite, which are all part of a major post-Yavapai/pre-Mazatzal orogeny regional quartzite succession (Figure 1).

THE PROTEROZOIC ROCKS OF THE UNCOMPAHGRE PLATEAU

The Paleoproterozoic rocks of the Uncompahgre Plateau (Figure 2) include sillimanite-bearing, quartzofeldspathic migmaitite (1,741 ± 11 Ma) [Livaccari et al., 2001] and a foliated meta-monzogranite pluton (1,721 ± 14 Ma) [Case, 1966; Livaccari et al., 2001]. Ambiguous zircon morphology leaves it unclear whether the 1,741 Ma date for the migmaitic gneiss represents the age of migmatization or the protolith age [Case, 1966]. Investigations by Livaccari et al., [2001] reveal that the migmaitic gneiss has a pervasive S1 foliation that is folded into tight to isoclinal F2 folds that plunge northwest (Figure 2). NE-trending F3 folds overprint the NW-trending F2 folds, and the resulting interference pattern is that of vertically elongated basin and dome structures (Figure 2). The monzogranite pluton intrudes migmaitic gneiss and contains a fabric defined by flattened xenoliths and aligned alkali-feldspar phenocrysts and biotite. This monzogranite pluton is interpreted as syn-deformational with the final phases of D2 shortening [Livaccari et al., 2001]. Post-kinematic intrusive rocks include the megacrystic Vernal Mesa monzogranite (1,430 ± 22 Ma) [Bickford and Cudzilo, 1975], rare NW-SE-striking, biotite-hornblende dikes (1,366 Ma 40Ar/39Ar biotite), and tourmaline-bearing pegmatite dikes (Figure 2) [Livaccari et al., 2001].

U-Pb GEOCHRONOLOGY

U-Pb zircon/titanite geochronology was employed to constrain the crystallization age of the Vernal Mesa monzogranite. Zircon and titanite fractions were hand picked, examined using a petrographic microscope, characterized by cathodoluminescence, extensively abraded, and then subjected to a final optical re-evaluation before analysis. For complete methods please see appendix of Connelly and Mengel [2000].

The Vernal Mesa Monzogranite: Sample K00BC9.

We chose to re-date the Vernal Mesa monzogranite because the existing Rb-Sr date of 1,480 ± 40 Ma was considered unreliable and because the pluton’s syn-kinematic nature provides a timing constraint on the reactivation of the Black Canyon shear zone. The Vernal Mesa pluton is a very coarse-grained, porphyritic, weakly-to-strongly-foliated monzogran-
Plate 2. Compositional maps used for in situ monazite geochronology on two monazite grains from sample MJBC43. Colors denote relative abundances of Y, Th, Ca and U. There is very little chemical zoning in either. a) Monazite inclusion in a cordierite porphyroblast that yields an age of 1,390 ± 6.3 Ma. b) Monazite inclusion in a garnet porphyroblast that yields an age of 1,389.8 ± 5.9 Ma. There are no older cores in either of these minerals.

A) Inclusion in corderite 1,388.8 +/-6.3 Ma

B) Inclusion in garnet 1,389.8 +/-5.9 Ma
ite to granodiorite. Average normative composition of 4 samples is: 43% plagioclase, 22% orthoclase, and 15% quartz; other modal minerals are biotite, opaque iron minerals, epidote, titanite, hornblende, apatite, and calcite [Hansen, 1971]. Three zircon fractions (Z2, Z4, and Z5) and one titanite fraction (T1; not used in calculation) are slightly discordant but all have similar 207Pb/206Pb ages that average to 1,434 ± 2 Ma (Figure 4; Table 1, see CD ROM in back cover sleeve). We interpret the crystallization age as 1,434 ± 2 Ma based on a weighted mean of these points (Figure 6). The ca. 1,409 Ma titanite suggests that the area either remained hot for tens of millions of years or was reheated ~ 20 million years after pluton emplacement. Zircon fractions Z1 and Z3 are interpreted to contain an inherited component.

**IN SITU MONAZITE GEOCHRONOLOGY**

We conducted *in situ* monazite geochronology to determine the timing of metamorphic events as recorded in the host rock bounding the Vernal Mesa monzogranite (Plate 2). The compositional mapping, age mapping, and microprobe dating of *in situ* monazite is a powerful tool that enables complex tectonic histories to be integrated [Williams et al., 1999]. Monazite is a rare earth element-bearing (dominantly cerium) phosphate that also contains Th, U, and insignificant levels of nonradiogenic lead and is therefore widely used for U-Pb dating [Parrish, 1990]. Because diffusion of major and trace elements is slow, monazite retains geochronological data through younger metamorphic events [Williams et al. 1999]. The concentrations of U, Th and Pb are used to calculate an age assuming a negligible amount of common Pb and no significant modification of their concentrations by mass transfer [Williams et al., 1999]. Elemental concentrations are measured *in situ* on the electron microprobe and thereby provide a unique opportunity to infer the age of deformatonal fabrics.

Sample MJBC43 is from a schist ~3 km southeast of the Vernal Mesa monzogranite (Plate 1). This rock contains porphyroblasts of garnet-staurolite-cordierite-anthophyllite surrounded by a matrix of plagioclase and quartz. The pelite grid for the KFMASH system and the empirical petrogenetic grid for cordierite-anthophyllite rocks define an overlapping stability field of ~600 ± 50°C and 3 ± 1 kbar [Spear, 1995]. Two monazite minerals were dated using *in situ* monazite geochronology. One is an inclusion in a cordierite porphyroblast that yields an age of 1,390 ± 6.3 Ma (Plate 2a); the second is an inclusion in a garnet porphyroblast that yields a similar age of 1,390 ± 5.9 Ma (Plate 2b). Age zoning is absent within both of these grains. Because these monazites are inclusions within cordierite and garnet porphyroblasts, we interpret them as a timing constraint on the growth of the metamorphic assemblage and an indication that metamorphism outlasted emplacement of the Vernal Mesa pluton during Mesoproterozoic intracratonic tectonism in the Black Canyon shear zone.

**REGIONAL IMPORTANCE OF NW-TRENDING FABRICS IN SOUTHWESTERN COLORADO**

From the Cochetopa succession in the east to the Proterozoic rocks of the Uncompahgre Plateau on the west, the dominant regional foliation trends northwest (ca. 1.72–1.70 Ga). In the ~40 km wide region that bounds the NE-striking Black Canyon shear zone, F2 folds and L2 stretching lineations are reoriented to north-south trends. We relate this progressive reorientation of F2 fold axes to movement on the NE-striking Black Canyon shear zone. The foliation trajectories of this transition suggest that the Green Mountain and Coffee Pot Hill antiforms (F2) are macro-scale drag folds that were rotated into their NE-trend during dextral shear (Plate 1). To explain this observation, we propose that the early assembly of this portion of the Yavapai province was along NW-striking features that developed ca. 1.72–1.70 Ga, not the NE-striking orogenic fabric that is dominant throughout much of the Yavapai province.

To determine the regional extent and depth of these NW-striking features we examined aeromagnetic and mantle tomographic trends. The aeromagnetic map for the central U.S.
**Plate 3.** a) The Magnetic Anomaly Map of North America for southern Wyoming, Colorado, and northern New Mexico \([\textit{NAMAG}, 2002]\). Relative differences in magnetic susceptibility of sub-sediment rocks are shown as various shades between pink (highest) and dark blue (lowest). Notice the location of the Vernal Mesa monzogranite (white circle) that appears as a NE-striking red zone that crosses a NW-striking blue zone. b) P-wave velocity map for Colorado and New Mexico from \textit{Dueker et al.}, [2001]. The warm colors (red/yellow) show regions of mantle at ~100 km depth that have slow velocities relative to the cold colors (green/blue). The yellow/green contour represents relatively low velocity zones and is shown as a green line in the compilation map. c) Compilation map of low velocity zones, low magnetic susceptibility areas, and ca. 1.75 – 1.65 Ga Paleoproterozoic geology. There is a general coincidence between surface fabrics, aeromagnetic and tomographic anomalies along the NE-striking Cheyenne belt and Jemez zone as well as the NW-striking Uncompahgre/Gunnison zone. This agreement in data sets suggests the presence of a NW-trending structure from the surface to ~100 km.
contains several prominent NE-and NW-striking zones of magnetic highs and lows that record contrasting magnetic characteristics of Proterozoic basement rocks within the upper few meters of basement (Plate 3a) [NAMAG, 2002]. The Vernal Mesa monzogranite appears as a ~6-km-long NE-striking zone of high magnetic susceptibility in the Black Canyon area that crosses a NW-striking zone of low magnetic susceptibility (Plate 3a). Two NE-striking zones of magnetic lows coincide with the Cheyenne belt and Jemez lineament, notable major Proterozoic features [Karlstrom and Humphreys, 1998]. The Proterozoic rocks of the Cochetopa, Dubois, Black Canyon, and Uncompahgre Plateau areas coincide with a zone of NW-striking magnetic lows, which is consistent with field observations (Plate 3c), and suggests that the NW-striking fabric is a regionally significant crustal-scale feature.

Seismic tomography of the western U.S. shows zones of low velocity at 100-km-depth that are interpreted as sites of partial melting of hydrated olivine-poor lithosphere, such as oceanic crust [Deuker et al., 2001]. The “warm zones” of crustal melting thus reveal older hydrated compositional domains (Plate 3b) [Karlstrom and CD-ROM working group, 2002]. The low velocity zones coincide with several NE-striking features that are proposed as Paleoproterozoic suture zones (e.g. the Saint George and Jemez lineaments) [Deuker et al., 2001]. From these results it is proposed that young lithospheric melting has been preferentially focused along Paleoproterozoic suture zones [Deuker et al., 2001; Karlstrom and CD-ROM working group, 2002]. The one exception to the dominant NE-striking trend is the Aspen anomaly, a low-velocity body that, although not completely defined, appears to have a NW-striking aspect under southwestern Colorado (Plate 3b) [Deuker et al., 2001]. By comparison with the NE-striking low-velocity lineaments interpreted as sites of preferential melting associated with fossil suture zones, the NW-striking Aspen anomaly may define a suture zone of contrasting orientation, or at least imply that NW-striking elements are present in both the crust and upper mantle beneath southwestern Colorado (Plate 3c).

PLATETECTONIC MODEL FOR CRUSTAL ASSEMBLY

A tectonic model for the assembly of the Yavapai province to the Archean craton must explain: (1) disparate ages of the volcanogenic sequences across the region that are interpreted to be components of arcs: Irving formation (1.80–1.78 Ga) [Gonzales, 1994], Dubois succession (1.77–1.76 Ga) [Bickford et al., 1989], Cochetopa succession (1.74–1.73 Ga) [Bickford et al., 1989], Green Mountain (1.78–1.76 Ga) [Tyson et al., 2002], and the Granite Gorge Metamorphic Suite (1.75–1.73 Ga) [Ilg et al., 1996; Hawkins et al., 1996]; (2) data on the timing of multiple phases of deformation of these arcs with the northeast grain created ~1.70–1.68 Ga and northwest grain active earlier at ca. 1.72–1.71 Ga; (3) evidence that NEn-trending fabrics were forming along the Cheyenne belt at 1.78–1.75 Ga coeval with the deposition of volcanogenic rocks that assembled across NW-striking structures in southwestern Colorado; and (4) the presence of older fragments of continental crust such as the Elves Chasm gneiss (1.84 Ga) [Hawkins et al., 1996; Ilg et al., 1996] and as evidenced by inherited zircons 1.87–1.84 Ga [Hill and Bickford, 2001].

The tectonic model presented here identifies analogous tectonic elements within the modern collision of the Australian continental plate margin with the Indonesian system (Figure 7a) [Hamilton, 1979] and proposes that a comparably complex margin existed along Laurentia during the Paleoproterozoic. The Banda arc (shown “upside-down” in Figure 7a) is an active subduction system composed of an inner and outer arc that trends eastward (left) from Java for ~2,000 km then turns nearly 180° (Figure 7a) [Hamilton, 1979]. Upper Cenozoic calc-alkaline volcanic rocks dominate the inner ridge from Bali to Wetar, whereas the outer ridge is composed of Tertiary subduction melange and imbricated complexes in Timor [Hamilton, 1979]. The outer ridge islands are subduction-imbricated complexes composed of continental crust, while the inner arc islands are formed by the convergence and ramping of material during the advance of island arcs onto continental crust [Hamilton, 1979]. The model proposed for this complex geometry involves: (1) the creation of the inner arc along a continuous subduction system with only a slight arcuate geometry, and (2) counterclockwise rotation of the subduction system as shortening progressed [Hamilton, 1979]. Other features that are important to note are arcs that are constructed on fragments of older continental crust (e.g., Sumba) and oceanic crust as well as the presence of transform faults.

A possibly analogous geometry is proposed for the assembly of the Yavapai province and accretion to the southwestern margin of Laurentia. The model for the assembly of the northernmost Yavapai province along NE-striking structures may have begun with a tectonically thinned post-2.1 Ga rifted Archean craton [Karlstrom and Houston, 1984; Tyson et al., 2002]. Ocean basin closure proceeded via the south-dipping subduction system that built the Green Mountain arc (Figure 7b; 1.78–1.76 Ga) [Tyson et al., 2002]. The continental margin reached the Green Mountain trench between 1.77–1.76 Ga, and terminated south-dipping subduction. The Rawah arc (1.78–1.72 Ga) then approached from the south over a south-dipping subduction zone and collided with the Proterozoic continental margin between 1.74–1.73 Ga [Tyson et al., 2002].

Using the Banda arc analog, nearly coeval arc sequences that are presently located southwest of the Cheyenne belt are hypothesized to have been built along a subduction zone with
varying orientation (northwest to north-south). Convergence between the arcs and Archean craton continued, possibly with a component of counterclockwise rotation of the subduction zone and arc into a NW-trend. The Granite Gorge metamorphic suite (1.75–1.73 Ga) is an example of a section of arc that was built partially on top of a fragment of older continental crust, the Elves Chasm gneiss (Figure 7b; 1.84 Ga) [Hawkins et al., 1996; Ilg et al., 1996]. The inherited zircons from the Dubois and Cochetopa successions may also record presently unexposed older continental fragments under southern Colorado [Hill and Bickford, 2001]. Transform faults may have dissected the ocean crust between arcs to accommodate rota-

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**Figure 7.** a) A simplified geologic map of the Banda Sea that is inverted (north-down) to show the current geometries of orthogonal subduction related to assembly of island arcs onto the northern cratonic margin of Australia [Hamilton, 1979]. b) The proposed distribution of major island arcs in the southwestern U.S. that formed during the Paleoproterozoic. The hypothetical geography demonstrates the possibility of a complex interaction between NW- and NE-striking subduction zones ca. 1.75 Ga that will presumably create fabrics in the middle crust during assembly of the Yavapai province between 1.73–1.70 Ga.
tion, such as the Buckhorn Creek shear zone [Cavosie and Selverstone, in press]. The Mojave province was also built to the southwest possibly above a NW-striking subduction zone as proposed by Duebendorfer et al. [2001]. The arc terranes were then assembled along NW-striking structures to form the Yavapai province between 1.73–1.71 Ga. (Figure 7b; see CDROM in back cover sleeve) [Albin and Karlstrom, 1991; Duebendorfer et al., 2001; Lafrance and John, 2001; Livaccari et al., 2001]. The NW-striking structures were later overprinted (rotated and tightened) during the assembly of the Yavapai province onto the southern margin of Laurentia (1.70–1.68 Ga). This preliminary model considers evidence from Paleoproterozoic rocks (inherited zircons, transform faults, suture zones, arcs, fragments of older continental crust, assembly along NE and NW-striking boundaries) and attempts to assemble this evidence into a unifying model based on a modern tectonic setting, the Banda Sea.

**TECTONIC OVERVIEW**

Combining data from the Black Canyon area with that from surrounding exposures of Proterozoic rocks, we suggest that the evolution of continental crust in the region involved six main tectonic/plutonic stages:

1) **Deposition of a bimodal volcanic succession and related turbidite packages (ca 1.79–1.73 Ga).**

This stage involved: (1) Submarine volcanism (1.77–1.76 Ga) in the Dubois succession and intrusion of cale-alkaline granites (1.757–1.751 Ga) [Bickford et al., 1989; Hill and Bickford, 2001]; (2) Deposition of the 30 m.y. younger, submarine Cochetopa succession in the same region (1.74–1.73 Ga) [Bickford et al., 1989; Hill and Bickford, 2001]; (3) Deposition of the Black Canyon (>1.74 Ga) [Jessup et al., 2002b] and the Uncompahgre (1.741 Ga) metasedimentary successions [Livaccari et al., 2001]; (4) Early isoclinal F1 folding and creation of pervasive S1 foliation. The long duration of arc-type magmatism and related deposition of immature sediments suggest one or more long-lived oceanic subduction systems as depicted in Figure 7b.

2) **Outboard assembly stage (1.73–1.70 Ga).**

This stage involved the formation of numerous NW–trending, variably plunging F2 folds throughout the Uncompahgre, Black Canyon, Dubois and Cochetopa successions at ca. 1.70–1.72 Ga, presumably during assembly of arcs along NW-striking boundaries. Timing of NE-SW directed shortening to form NW-trending folds is constrained to be ~1.72–1.70 based on: 1) the deposition of the pre-D2 Cochetopa succession (1.74–1.72 Ga) and late-syn-NE-directed shortening in the tonalite of Gold Basin (1.713 Ga) [Afifi, 1981; Wortman and Bickford, 1990; Lafrance and John, 2001]; 2) the emplacement of the Granite of South Beaver Creek at ca. 1.70 Ga (outer ring of the Gunnison annular complex) syn-NE-directed shortening [Lafrance and John, 2001; Hill and Bickford, 2001]; and 3) the syn-NE-directed shortening emplacement of the megacrystic monzogranite in the Uncompahgre Plateau (1.721 Ga) [Livaccari et al., 2001].

3) **Assembly to Laurentia stage.**

This stage involved the rotation, tightening and transposition of NW- to NS-trending F2 folds into isoclinal F3 folds with a strong subvertical axial planar S3 foliation in the Black Canyon shear zone during dextral shear. L2 lineations and F2 fold axes were rotated towards vertical in the highest strain zone. Strain was partitioned throughout the shear zone resulting in various degrees of fabric intensity and transposition. In the Uncompahgre Plateau, basin and dome interference patterns were created during D3 by NE-trending F3 folds that overprint the NW-trending F2 folds [Livaccari et al., 2001]. NE-trending F3 folds in the Cochetopa and Dubois successions fold the southern end of the Iris syncline, South Beaver Creek, Gold Basin antiforms [Afifi, 1980; Shonk, 1984]. Although not yet well dated in the Black Canyon region, F3 is constrained as ~1.68 Ga by monazite in the Homestake shear zone [Shaw et al., 2001].

4) **Quartzite deposition (ca. 1.70–1.68 Ga).**

Clean quartzites exposed in the Cebolla Creek syncline and Needle Mountains [Gibson and Harris, 1992] were unconformably deposited onto the previously deformed Paleoproterozoic basement. The presence of clean quartzites suggests a transition from oceanic arc-related deposition to a more stable continental setting before the onset of the Mazatzal orogeny.

5) **The Mazatzal orogeny (ca. 1.65 Ga).**

This stage involved the assembly of a younger tectonic block to the southeast margin of the Yavapai province along a complex, northeast-striking boundary [Shaw and Karlstrom, 1999]. NW-SE directed shortening rotated NW-trending fabrics into a northeast orientation and thereby reactivated and/or created NE-striking shear zones that overprinted the NW-striking F2 folds. The NW-directed shortening also folded quartzites into NE-striking synclinal keels (Coal Creek, Blue Ridge, Cebolla Creek, Uncompahgre Group).
6) Intra-continental tectonism (ca. 1.43–1.3 Ga).

The emplacement of the Vernal Mesa monzogranite in the Black Canyon (1,434 ± 2 Ma) took place during predominantly dextral, southeast side down transpressional shear parallel to the preexisting Black Canyon shear zone (S3). Based on metamorphic assemblages and in situ monazite geochronology, the schist ~3 km from the Vernal Mesa monzogranite experienced metamorphism at a pressure and temperature of 600 ± 50°C and 3 ± 1 kbar at 1.39 Ga.

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