Tectonic inferences from the ca. 1255–1100 Ma Unkar Group and Nankoweap Formation, Grand Canyon: Intracratonic deformation and basin formation during protracted Grenville orogenesis

J. Michael Timmons†
New Mexico Bureau of Geology and Mineral Resources, New Mexico Tech, Socorro, New Mexico 87801, USA

Karl E. Karlstrom‡
Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, New Mexico 87131, USA

Matthew. T. Heizler§
New Mexico Bureau of Geology and Mineral Resources, New Mexico Tech, Socorro, New Mexico 87801, USA

Samuel A. Bowring⁶
Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts 02139, USA

George E. Gehrels⁷
Department of Geosciences, University of Arizona, Tucson, Arizona 85721, USA

Laura J. Crossey‡‡
Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, New Mexico 87131, USA

ABSTRACT

The Unkar Group is one of the best-preserved remnants of Mesoproterozoic sedimentary rocks in the southwestern United States. It provides an exceptional record of intracratonic basin formation and associated tectonics kinematically compatible with protracted “Grenville-age” NW-directed shortening. New U/Pb age determinations from an airfall tephra at the base of the Unkar Group dates the onset of deposition at ca. 1255 Ma, and ⁴⁰Ar/³⁹Ar K-feldspar thermochronology in the Grand Canyon indicates that basement rocks cooled through 150 °C between ca. 1300 and 1250 Ma, refining exhumation rates of basement rocks just prior to Unkar deposition. Abrupt thickness and facies changes in conglomerate and dolomite of the Bass Formation (lower Unkar Group) associated with NE-striking monoclines flexures indicate NW-directed synsedimentary contraction at ca. 1250 Ma. A large disconformity (~75 m.y. duration) is inferred between the lower and upper Unkar Group and is located below the upper Hakatai Shale, as documented by detrital zircons. A second style of Unkar Group deformation involved the development of half grabens and full grabens that record NE-SW extension on NW-striking, high-angle normal faults. Several observations indicate that NW-striking normal faulting was concurrent with upper Unkar deposition, mafic magmatism, and early Nankoweap deposition: (1) intraformational faulting in the Bass Formation, (2) intraformational faulting in the 1070 Ma (old Rb/Sr date) Cardenas Basalt and lower Nankoweap Formation, (3) syntectonic relationships between Dox deposition and 1104 Ma (new Ar/Ar date) diabase intrusion, and (4) an angular unconformity between Unkar Group and Nankoweap strata. The two tectonic phases affecting the Unkar Group (ca. 1250 Ma and ca. 1100 Ma) provide new insight into tectonics of southern Laurentia: (1) Laramide-style (monoclines) deformation in the continental interior at ca. 1250 Ma records Grenville-age shortening; and (2) ca. 1100 Ma detrital muscovite (Ar/Ar) and zircon (U/Pb) indicate an Unkar Group source in the Grenville-age highlands of southwestern Laurentia during development of NW-striking extensional basins. We conclude that far-field stresses related to Grenville-age orogenesis (NW shortening and orthogonal NE-SW extension) dominated the sedimentary and tectonic regime of southwestern Laurentia from 1250 to 1100 Ma.

Keywords: Grenville, Unkar Group, Nankoweap Formation, Precambrian monocline, intracratonic rifting, Grand Canyon.

INTRODUCTION

The late Mesoproterozoic (1.3–1.0 Ga) was characterized by the development of orogenic belts worldwide that record the assembly of the supercontinent of Rodinia (Dalziel, 1991; Hoffman, 1991; Moores, 1991). The Grenville orogen of NE Laurentia (Rivers et al., 2002) and the Texas Grenville (Mosher, 1998; Bickford et al., 2000) record protracted convergence along the “southern” (present coordinates) plate margin. The term “Grenville” has been used in many ways, but here, we consider a broad interval of Grenville-age orogenesis (1360–1232 Ma), and continent-continent (1150–1120 Ma) collisions (Mosher, 1998;
This protracted deformation was the culmination of a long-lived convergent/transpressional margin that persisted between 1.8 and 1.0 Ga (Karlstrom et al., 1999, 2001).

Modern examples of far-field intracratonic deformation (interior to orogenic belts) in response to plate-margin interaction are common and include Lake Baikal, Tibetan Plateau, and upper and lower Rhine grabens (discussed in the following). In each of these cases, deformation in the continental interior is driven by and is kinematically linked with plate-margin deformation. Ancient examples of far-field intracratonic deformation in southwestern Laurentia associated with plate-margin interaction have been proposed for several time periods, such that this region seems to be an exceptionally important laboratory for understanding intracratonic deformation. Important examples include: regional 1.4 Ga magmatism and deformation (Nyman et al., 1994; Kirby et al., 1995); 1.1 Ga intracratonic rift in the midcontinent rift (Gordon and Hempton, 1986) and Central Basin platform (Adams and Keller, 1996); ca. 750 rifting of the continental interior ~200 km from the plate margin (Timmons et al., 2001); Ancestral Rocky Mountains (Kluth and Coney, 1981; Ye et al., 1996; McBride and Nelson, 1999); and Laramide orogeny (Hamilton, 1981; Bird, 1988; Livaccari, 1991; Erslev and Rogers, 1993; Humphreys, 1995).

This paper examines the stratigraphic and structural record of intracratonic contractional and extensional tectonism and syntectonic sedimentation in the 1.25–1.1 Ga Unkar Group and ca. 900 Ma Nankoweap Formation of the Grand Canyon. We recognize a complex interplay between synsedimentary contractional and extensional faults in the Unkar Group, which demonstrates a kinematic link to Grenville orogenesis (Fig. 1). We present new geo-chronology that improves our understanding of regional tectonic events and sedimentary patterns. Finally, this paper illustrates contrasting responses to plate-scale contractional deformation from proximal deposits in the foreland to distal deposits hundreds of kilometers into the continental interior. Combined with analysis of correlative units of the Apache Group of Arizona (Shrid, 1967; Wurcke, 1989), Grenville deposits in west Texas, and the Lower Pah Rump Group of California (Prave, 1998), this study leads to a new paleogeographic understanding of southwestern Laurentia, which may be useful for continued tests of conflicting Rodinia reconstructions (Brookfield, 1993; Dalziel, 1997; Karlstrom et al., 1999; Piper and Jiasheng, 1999; Sears and Price, 2000; Wingate et al., 2002).

**BACKGROUND OF GRAND CANYON GEOLOGY**

The Grand Canyon Supergroup is exposed as isolated fault-bounded remnants along the main stem of the Colorado River and its tributaries in the Grand Canyon (Fig. 2). It rests nonconformably on basement metamorphic and igneous rocks of the Granite Gorge Metamorphic Suite (Ilg et al., 1996). The Grand Canyon Supergroup is formally divided into the Upper Mesoproterozoic (1255–1100 Ma) Unkar Group and Neoproterozoic (ca. 800–742 Ma) Chuar Group (Van Gundy, 1951). The unconformity-bounded Nankoweap Formation separates the two groups. The Sixtymile Formation caps the Chuar Group, and all Proterozoic rocks are overlain in angular discordance (up to 15°) by Middle Cambrian Tapeats Sandstone.

The 1255–1100 Ma Unkar Group is ~2100 m thick and is divided into the Bass Formation, Hakatai Shale, Shinumo Sandstone, Dox Formation, and Cardenas Basalt (Fig. 3; Elston, 1979; Hendricks and Stevenson, 1990). The succession contains both fluvial and shallow-marine deposits, with one main disconformity within or below the Hakatai. In general, Unkar rocks dip northeast (10–30°) toward normal faults that dip 60° southwest (Sears, 1990).

John Wesley Powell (1875) first described the gently tilted strata along the river corridor. Walcott (1894) named the upper Chuar and lower Unkar terranes. Noble (1914) divided the Unkar Group into five formations, excluding the Cardenas Basalt, but the stratigraphy was later revised by Beus et al. (1974) to include the Cardenas Lavas within the Unkar Group. The use of the term Cardenas Lava(s) is well engrained in the literature (Hendricks and Stevenson, 1990, 2003); however, we use the term Cardenas Basalt after Larson et al. (Larson et al., 1994) as the preferred name for the voluminous basalt and basaltic andesite at the top of the Unkar Group. Noble (1914), Sears (1973), and Timmons et al. (2001, 2003) mapped the structures in the Unkar Group. Numerous workers conducted further stratigraphic studies of the formations of the Unkar Group (Beus et al., 1974), including unpublished MS theses that are summarized and updated in the following text (Fig. 3).

**SYNOPSIS OF UNKAR GROUP STRATIGRAPHY**

Dalton (1972) described the Hotauta Conglomerate and Bass Limestone. He recognized the heterolithic composition of the Bass Limestone, and suggested that it should have formation status, including the Hotauta Conglomerate as a member. Dolomite is the dominate carbonate facies in the Bass Formation, with subordinate conglomerate, breccia, sandstone, and mudstone (60–100 m; Fig. 4). These intercalations and primary structures, such as wave rippled sandstone and mud-cracked surfaces, indicate that Bass Formation deposition occurred during repeated subaerial exposure and flooding, and that it represents relatively low-energy intertidal to supratidal depositional environments in a general transgressive sequence (Dalton, 1972; Beus et al., 1974; Hendricks and Stevenson, 1990). White, very fine-grained tephra are interbedded with dolomite and mudstone toward the base of the section. One of these ash fall beds yielded zircons for U/Pb geochronology (see geochronology section).

The Hotauta Member conglomerate at the base of the Bass Formation in the eastern Grand Canyon contains more than 80% clasts of granite and quartzite that range from pebble to cobble size (Table 1). Quartzite clasts have no local equivalents in the Grand Canyon, indicating a distant source. Granite clasts may be locally derived and mixed with quartzite clasts, or transported with quartzite clasts from a distant source. Conglomerate beds and intraclastic breccias indicative of higher-energy depositional environments are interbedded with low-energy carbonate and mudstone deposition within the Bass Formation. Clast composition and size in the upper conglomeratic units are similar to the basal Hotauta Member, but with the addition of intraclasts of carbonate and siliciclastic composition derived from the lower Bass Formation (~17.3% of the conglomerate clasts, Table 1). Interbedded breccias are composed exclusively of intraclastic material and thus are interpreted to be locally derived (Fig. 4). The recycling of carbonate, and perhaps clasts from basal conglomerates in the eastern Grand Canyon, reflects localized uplift (by faulting) and erosion of lower units.

The Hakatai Shale (137–300 m) is a mudstone- to coarse sandstone–dominated package that appears to be in gradational contact with the Bass Formation (Reed, 1976). The Hakatai Shale has been subdivided informally into three members. The Hance Rapids member is dominated by thin-bedded subarkose to quartz arenite. The middle member (Cheops Pyramid) is mostly mudstone. The uppermost Stone Creek member is dominated by coarse arkose. The Hakatai Shale is more heterolithic than the name implies, because it contains numerous sandstone beds. Marker beds generally are absent in the section; however, two prominent sandstone markers (~0.5 m thick) are present ~50 m above the base. Mud cracks, ripple marks, tabular-planar cross-bedding, salt casts,
Figure 1. 1250–1100 Ma intracratonic structures and basins: 1.25–1.1 Ga sedimentary basins (gray), tectonic elements, and magmatic rocks (black) provide a record of assembly of Rodinia and inboard stresses related to Grenville convergence. Light gray represents hypothesized area of inferred intracratonic seaway at ca. 1250 Ma, based on lithologic correlation of age-equivalent carbonate sequences in the Southwest (Unkar, Apache, Castner, and Allamoore; located by stars). Inset shows alternate Rodinia reconstructions: AUSMEX reconstruction (Wingate et al., 2002), AUSWUS (Brookfield, 1993; Karlstrom et al., 1999; Burrett and Berry, 2000), and SWEAT (Moores, 1991). All ages are in billions of years. Map modified from Timmons et al. (2001).
and tool marks, indicate shallow-water deposition, probably in a marginal-marine/tidal-flat environment (Reed, 1976).

The contact of the Hakatai Shale with the overlying Shinumo Sandstone is sharp and locally truncates cross-beds and channel forms, which suggests a hiatus (Fig. 3; Daneker, 1975). The Shinumo Sandstone (syn. Shinumo Quartzite; 355–410 m) forms resistant cliffs composed of lower subarkose and upper quartz arenite with subordinate interbedded mudstone (Daneker, 1975). A basal lag of conglomerate contains basement clasts up to 5 cm across. Like the Hotauta Conglomerate, the basal conglomerate includes quartzite clasts that have no known equivalents in the Grand Canyon region.

The Shinumo Sandstone is divided informally into the basal Surprise Valley, Ribbon Falls, Papago Creek, Cottonwood Camp, and 75-Mile Rapid members from bottom to top. The predominant rock type in the Shinumo Sandstone is quartz arenite; however, subarkose is more common in the Surprise Valley member. Sedimentary structures in sandstone beds are dominated by centimeter- to meter-scale planar-tabular cross-stratification and trough cross-beds (Fig. 5) that record north-directed paleocurrent directions. Subordinate bidirectional paleocurrent indicators, such as bidirectional cross-bed sets, are also observed. Trough cross-beds suggest a more northerly transport direction and are more common near the top of the section and its contact with the Dox Formation. Upper massive sandstone beds of the Shinumo contain dramatic convolute bedding. The presence, abundance, and repetition of very thick (meters to tens of meters) contorted beds in upper beds have been cited as evidence for earthquake activity and fluid migration during Shinumo deposition (Sears, 1973; Daneker, 1975; Middleton and Blakey, 1998; Timmons et al., 2001). The depositional environment proposed for the Shinumo is nearshore, marginal marine, and/or fluvial/deltaic (Daneker, 1975). Sears (1973) reports that faulting during Shinumo deposition resulted in thickness changes across discrete fault zones. This conclusion was based on mapping in Bright Angel Canyon and a reported thickness change of ~61 m over 460 m distance across the Bright Angel monocline.

The contact between the Shinumo Sandstone and the Dox Formation is gradational and is marked by a change in topographic expression and color. The transition also marks a striking upward gradation from quartz arenite to mudstone and fine-grained arkose. The basal Dox Formation includes channel facies arkosic sandstone that represents fluvial deposition by a large river system, and hence a regression to fluvial/deltaic facies.

The Dox Formation is divided into the Escalante Creek, Solomon Temple, Comanche Point, and Ochoa Point Members, based
on color changes and topographic expression (Stevenson, 1973). Marker beds generally are absent with the exception of a zone of contorted bedding near the base that may suggest a continuation of Shinumo-age seismicity, and two beds in the Comanche Point Member, a convolute sandstone bed and a thin (30 cm) carbonate bed (Stevenson, 1973).

The Dox Formation is interpreted here as a marginal marine deltaic to tidal flat sequence that records a regression from marine Shinumo followed by transgression within the Dox Formation. Fluvial-dominated distributary channel deposits with unidirectional paleocurrent indicators dominate the base of the Dox section (Escalante Creek Member; Fig. 5). Upper members of the Dox Formation record the transition from channel sandstone beds in intrafluvial mudstones to thin-bedded sheet sand deposits with wave-ripples that are mud-drapped and mud-cracked. Uppermost beds are interpreted as tidal-dominated mud-flat facies.

The contact between the Dox Formation and Cardenas Basalt shows interfingering of basalt flows with Ochoa Point sandstone and mudstone. Small folds and load structures beneath the basalt, and intraflow red-bed sandstone beds through the Cardenas section (Hendricks, 1972; Stevenson, 1973) suggest that basaltic volcanism was contemporaneous with red-bed deposition.

The Cardenas Basalt is ~300 m thick and consists of >10-m-thick flow units (Hendricks, 1972). Three marker layers are described as informal members: the bottle-green member, fan-jointed member, and lapillite member (Lucchitta and Hendricks, 1983). Hyaloclastite of the bottle-green member is ~90 m thick, highly altered, and contains secondary chlorite, epidote, talc, and zeolites. Basaltic andesite (~50 m thick) comprises the fan-jointed member with porphyritic to aphanitic and vesicular textures (Hendricks and Lucchitta, 1974). The lapillite member ranges in thickness from a few meters to several tens of meters thick and is composed of scoriaceous lapilli and blocks (~10 cm) and volcanic bombs (<1 m) in matrix, suggesting proximity to a vent location (Lucchitta and Hendricks, 1983). The lapillite member is interbedded with massive flows of basalt that comprise the remaining thickness of the Cardenas Basalt.

Intrusive rocks of the Unkar Group are similar in mineralogy and chemistry to the basalts, suggesting that intrusive and extrusive rocks were coeval and shared a common source. Intrusive rocks occur as dikes and sills within the Unkar Group, with sills ranging in thickness from a few tens of meters to 300 m; dikes typically are much thinner and locally follow fault planes.
Figure 4. Measured sections of the Bass Formation in Grand Canyon, including unpublished measured sections from Dalton (1972), marked by a D. Some of the measured sections in this study repeat and confirm measured sections reported by Dalton (1972) and are highlighted by a T. Measured sections are presented using the contact of the Hakatai as a time datum. Measured sections show an overall thickening of the Bass Formation toward the west. Abrupt thickness and facies changes are observed in Vishnu Canyon and record synsedimentary development of the Vishnu monocline. Measured sections are keyed by numbers to Figure 2 (RM = river mile).
Extrusive rocks tend to be more silicic than the intrusive rocks, however, the lower bottle-green member is compositionally very similar to intrusive rocks, suggesting some differentiation of the parent magma after emplacement of the sills and eruption of lowermost flows (Lucchitta and Hendricks, 1983). Unkar Group igneous rocks are correlated to mafic intrusions of similar age in the southwest (Howard, 1991).

Van Gundy (1937, 1951) first recognized the unconformity-bounded section of quartzitic sandstones and mudstones above the Cardenas and named the strata the Nankoweap Group, removing it from the Unkar Group of Noble (1914). Mapping in the eastern Grand Canyon by Maxson (1961) introduced the name Nankoweap Formation. Later workers divided the Nankoweap into two informal members, including the lower ferruginous member and upper member separated by a disconformity with tens of meters of relief (Elston and Scott, 1976) separating rocks with greatly differing paleomagnetic directions (Elston and Scott, 1973). The Nankoweap type section in Basalt Canyon measures ~100 m thick (Gebel, 1978). The lower member is dominated by hematite-cemented sandstone and siltstone with lenses of lithic sandstone derived from the underlying Cardenas. The upper member is composed of siltstone and thin-bedded fine-grained red-bed sandstones toward the base and more massive meter-thick sandstone beds toward the top of the section. The capping white sandstone is composed of a fine-grained quartz-cemented quartz arenite. Abundant sedimentary features are identified in the section and include planar-tabular and trough cross-bedding, ripple marks, mud cracks, soft sediment deformation, and rare salt casts (Gebel, 1978). Combined, these features suggest that the Nankoweap Formation was deposited in a moderate- to low-energy, shallow-water environment, perhaps a structurally controlled marine or lake environment (Elston and Scott, 1976; Gebel, 1978).

**UNKAR GROUP THERMOCHRONOLOGY AND GEOCHRONOLOGY**

**U-Pb Geochronology of the Bass Formation**

The combination of a new U/Pb zircon date on an ash bed (direct date) and Ar/Ar dates on intrusive rocks (direct dates), with thermochronologic results from basement rocks and detrital grains (indirect dates) provides new constraints on the age of the Unkar Group. Samples for U/Pb zircon dating were collected from ash-fall deposits interbedded in the lower Bass Formation. Sample K7-77-7 is from a very fine-grained (clay-sized) white horizon with no obvious sedimentary features to suggest transport of material. The suspected tephra are a few to several tens of centimeters thick, and are located in the lower half of most measured sections of the lower Bass Formation. Sample K7-77-7 is from a very fine-grained (clay-sized) white horizon with no obvious sedimentary features to suggest transport of material. The suspected tephra are a few to several tens of centimeters thick, and are located in the lower half of most measured sections of the lower Bass Formation.
zircons were separated from sample K7-77-7, five of which were analyzed. Analytical details are presented in Table DR11 and Schmitz et al. (2003). Four of the five zircons cluster near concordia and one is discordant but yields a precise upper intercept of 1254.8 ± 1.6 Ma (Fig. 6). We recognize the possibility that these grains could represent inherited or detrital zircon grains based on the dominance of ca. 1250 Ma detrital zircons in the upper Unkar Group (discussed in the following). However, based on regional lithologic correlations (see discussion section), and the simple zircon population recovered, we interpret 1254.8 Ma as the depositional age of the lower Bass Formation.

Basement Thermochronology

The 40Ar/39Ar data from basement rocks generally are compatible with the tephra age and suggest exhumation prior to Bass deposition. Timmons et al. (2001) reported that the onset of Unkar Group deposition postdated ca. 1250 Ma and final closure to Ar loss in K-feldspar-bearing basement rocks. The 40Ar/39Ar studies of K-feldspars show that individual feldspars contain multiple diffusion domain sizes, and thus have variable closure temperatures or retentivity (150–325 °C) dependent on domain size within a single grain (Lovera et al., 1989; McDougall and Harrison, 1999). Because different diffusion domains in a K-feldspar are closed to Ar loss at different temperatures, single feldspar crystals can record a segment of the thermal history rather than a single point. Smaller diffusion domains release Ar at temperatures between 150 and 200 °C, whereas the largest diffusion domains remain retentive at temperatures approaching 325 °C.

Metamorphic and igneous feldspars were collected along the river corridor to evaluate basement thermal history in the Grand Canyon (Fig. 2). Twelve K-feldspars were analyzed and their age spectra are given in Figure 7 (Table DR2, see footnote one). The age spectra are complex and do not in all cases yield unambiguous age or thermal-history interpretations. Electron microprobe analyses of selected samples reveal variable degrees of alteration to sericite and slight albitization. K-feldspar intergrowths at scales ranging from submicron to tens of microns. We attribute general age spectra complexity to this alteration.

Many spectra reveal steep age gradients from ca. 600–1300 Ma during the initial 10% of 39Ar released, and steep gradients commonly are followed by an age decrease or undulatory pattern for the middle part of the spectra (e.g., Figs. 7A, 7B, and 7E). Following the age decrease, or saddle part of the spectra, the ages rise during the final ~50% of gas release. Due to these complexities, several samples cannot be used for multiple diffusion domain modeling (MDD method; Lovera et al., 1989). The causes for the initially old apparent ages followed by decreasing apparent ages are not well understood, and we suspect that it is caused primarily by 39Ar recoil and secondary alteration of the K-feldspars. This pattern spectrum also is generated by models that mimic recrystallization of large diffusion domains below their closure temperatures (Lovera et al., 2002). Aside from the overall age spectrum complexity, some samples (Figs. 7A, 7C, and 7I) have ages that are too young based on the 1255 Ma age from the Bass Formation. If the ages were meaningful, these samples would suggest that the basement was at elevated temperatures (>200 °C) during deposition of the Bass Formation. Given that these samples were collected below exposures of the Bass Formation, they could not have been at these temperatures during or after Bass deposition (Fig. 4). For these samples, the sericite alteration of the K-feldspars presumably causes apparent ages that are geologically too young with respect to the inferred post-Bass Formation thermal history.

In contrast to samples that appear too young and/or too complex, samples K7-115-3, T02-98-14, and T02-98-16 (Figs. 7F, 7K, and 7L) are considered to be well enough behaved to be used for MDD modeling (Fig. 8). Arrhenius plots (Figs. 8D–F) are constructed using the fraction of 39Ar released and the laboratory-heating schedule with the assumption of a plane-sheet diffusion geometry. Of the three samples chosen for MDD modeling, sample T02-98-14 has initial diffusion coefficients that yield a well-defined linear array, which indicates an activation energy (E) of 38.1 kcal/mol (Fig. 8E). The other spectra have poorly defined initial linear segments that probably are related to simultaneous degassing of several diffusion domains and/or slight alteration. We chose to model all the samples with E = 38.1 kcal/mol in order to facilitate sample comparison, but recognize that this is a poorly constrained assumption, since it has been shown that K-feldspars can have a range of activation energies (McDougall and Harrison, 1999). The poor resolution of E will primarily affect the absolute temperature of the thermal history and has very little effect on the cooling rates inferred from the models (Lovera et al., 1997). Log r/r0 plots for each sample (Fig. 8G–I) are constructed from the diffusion coefficients and reference Arrhenius Law (Lovera et al., 1991). Like nearly all...
Figure 7. Age spectrum and K/Ca diagrams for basement K-feldspars along the Colorado River corridor. Variably complex spectra are attributed to different thermal histories, complex K-feldspar/albite intergrowths, and degree of alteration to white mica. Horizontal line at 1255 Ma represents the age of the ash layer within the Bass Formation that unconformably overlies the basement. Intermediate parts of the age spectra that record ages younger than this are considered unreliable based on maximum post–Bass deposition basement temperatures (<200 °C) estimated from Supergroup stratigraphic thickness. Initial steep age gradients probably record heating to ~150 °C during the Neoproterozoic burial history of the basement. Samples are keyed to the map in Figure 2.
Figure 8. Multiple diffusion domain (MDD) results from the least complex basement K-feldspars. (A–C) Measured and model spectra. (D–F) Measured and model Arrhenius plots. (G–I) Measured and model $\log(D/r_o^2)$ plots. (J–L) Contours of time-temperature history paths output by MDD analysis. The activation energy ($E$) of the K-feldspars (D–F) is estimated from the initial linear segment recorded by T02-98-14 and the reference $D_o/r_o^2$ is determined by forcing the reference line through the diffusion coefficient recorded by the first heating step. Model thermal histories (J–L) indicate significant cooling between 1300 and 1225 Ma and also suggest variable Neoproterozoic temperature maxima between ~100 and 150 °C. Based on the overlying Bass Formation age of 1255 Ma (vertical line in J–L), the models predict cooling that is too late for samples K7-115-3 and T02-98-16. This perhaps represents poor intercalibration between Ar/Ar and U/Pb methods and/or inaccuracy of apparent Ar ages due to alteration of the K-feldspars. Sample T02-98-14 records a thermal history that is compatible with the Bass Formation age and indicates significant basement cooling, and presumably exhumation, between 1300 and 1270 Ma.
K-feldspars, the log \( r/r_p \) plots reveal significant differences in diffusion domain sizes and volume fractions and also provide a means to better visualize the kinetic parameters relative to the Arrhenius plots, because information about the amount of \(^{39}\)Ar represented by each diffusion coefficient can be represented on the log \( r/r_p \) diagram. Thermal histories (Fig. 8J–L) are determined following the methods of Quidelleur et al. (1997) and are constructed by forward modeling of the measured age spectrum within the temperature range defined by the kinetic data. Model age spectra are shown along with the measured spectra in Figure 8A–C.

The derived thermal histories share several characteristics, but most fundamentally record cooling from \(~250–300~\)°C to below \(150~\)°C between 1300 and 1225 Ma. For samples K7-115-3 and T02-98-16, the timing of cooling may be problematic with respect to the 1255 Ma zircon age on the ash layer within the Bass Formation. Provided the zircon data faithfully record the Bass Formation deposition, it is expected that the basement at the unconformity would have cooled prior to 1255 Ma. Two explanations for the timing problem could be poor intercalibration between the U/Pb and \(^{40}\)Ar/\(^{39}\)Ar methods (e.g., Renne et al., 1998; Min et al., 2000) or using K-feldspars that are unsuitable for MDD analysis. We have tried to deal with intercalibration issues by adopting the most recent values for the total \(^{40}\)K decay constant (5.4766 – 10/\(a\) [\(a\) is standard for annum or year]) and of Fish Canyon sanidine (28.27 Ma) by Kwon et al. (2002). These values still require significant inspection, but are designed to close the observed gap between U/Pb ages and \(^{40}\)Ar/\(^{39}\)Ar ages. The fine-scale age spectrum complexity of the chosen K-feldspars may suggest caution with respect to highly rigorous treatment of the thermal histories. We suggest that the K-feldspar data record significant cooling within about a 10–50 m.y. period prior to Bass deposition, but realize that this assertion requires further testing.

Another aspect of the K-feldspar data that may be geologically useful is the steep age gradients observed in the initial part of the age spectra (Fig. 7). The MDD analysis predicts post–1100 Ma temperatures for the basement K-feldspars between \(~100~\) and \(150~\)°C (Fig. 8I–L). Based on the burial history recorded by the supergroup sediments (\(~3.6~\) km), this temperature range would be expected. However, the timing of the maximum post–1100 Ma temperatures is variable and may reflect not only sediment accumulation, but also transient, perhaps fluid-driven, temperature excursions. More work is required to understand in detail the basement thermal history from 1255 Ma to deposition of the Cambrian Tapeats Sandstone.

**Cardenas Basalt Geochronology**

Final deposition of Unkar strata is marked by voluminous basaltic and andesitic flows of the Cardenas Basalt, which are chemically similar to and likely synchronous with diabase intrusions within the lower Unkar Group (Hendricks and Lucchitta, 1974) and Apache Group (Howard, 1991), Elston and McKee (1982) reported an Rb/Sr age for whole-rock samples of the Cardenas Basalt of 1070 ± 70 Ma. This age was refined by Larson et al. (1994) to 1103 ± 66 Ma based on additional Rb/Sr data. New Ar-Ar data from two diabase sills yield precise ages that may also better constrain the timing of Cardenas activity (Fig. 9; Table DR3 [see footnote one]). Both a hornblende and a biotite were dated from a diabase at Stone Creek (Fig. 9A–B). H98-131.6-2 biotite, which grew within the contact metamorphic assemblage below the sill, yields an overall flat age spectrum with initial ages slightly older than the plateau age of 1104 ± 2 Ma (Fig. 9A). The hornblende (H98-131.6-4), which is a phenocryst phase within the sill, has a similar age spectrum, but the apparent plateau age is significantly older at 1124 ± 2 Ma (Fig. 9B). Considering that the sill should cool quickly, we would expect these minerals to yield discordant apparent ages. Isochron analysis of the hornblende data indicates the possibility of excess \(^{40}\)Ar within the hornblende (Fig. 9C). Steps D–L define a linear array that suggests an age of 1115 ± 4 Ma and an initial trapped argon component with a \(^{40}\)Ar/\(^{39}\)Ar value of 470 ± 60, which is significantly higher than the atmospheric value of 295.5. This hornblende isochron age just overlaps with the biotite plateau age at 2\(\sigma\) uncertainty and appears to support an excess Ar problem for the hornblende. Our preferred interpretation for the Stone Creek sill age is given by the biotite result at 1104 ± 2 Ma. Biotite sample 98GC25 is from a sill located in Bass Canyon and yields a very complex age spectrum (Fig. 9D). The spectrum may be related to recoil redistribution of \(^{40}\)Ar during irradiation, and we suggest that the total gas age of 1114 ± 1 Ma may record the intrusion age. This interpretation by itself would be of limited value, however, the apparent age of ca. 1100 Ma is consistent with other sill ages and regional mafic magmatism at this time (Howard, 1991).

Interestingly, K/Ar and Ar/Ar dates for the Cardenas Basalt span a wide interval, from 700 to 1090 Ma, and were postulated to reflect cooling during movement on the Butte fault, coincident with deposition of the Sixtymile Formation and “Grand Canyon orogeny” (Elston, 1979; Elston and McKee, 1982). However, Larson et al. (1994) suggested that the range in K/Ar dates represents apparent ages and is an artifact of an alteration/heating event at low temperatures, perhaps related to Neoproterozoic riftling (Timmons et al., 2001). Hence the designation of a Grand Canyon “orogeny” should be discontinued and replaced by “Neoproterozoic (ca. 750 Ma) extensional episode” that includes all penecontemporaneous syntectonic deposits along western Laurentia (Young et al., 1979; Elston and McKee, 1982; Ross et al., 1989; Ross, 1991; Prave, 1999).

**Detrital Mineral Geochronology**

Geochronology of detrital zircon and muscovite grains provides another useful tool for evaluating the age of sedimentary units and yields some information about possible source terrains. As part of a large effort to conduct detrital mineral age analyses from several samples within the Grand Canyon Supergroup, we present a relevant subset of both \(^{40}\)Ar/\(^{39}\)Ar muscovite ages (Table DR4a, DR4b, see footnote one) and U/Pb zircon ages (Table DR5, see footnote one) from selected units. Total fusion and age spectrum plateau analyses from muscovite single crystals from the Escalante Creek Member of the Doye Formation (Fig. 10A) yield ages ranging from ca. 1120–1260 Ma, with a well-defined node at 1140 Ma. These young apparent ages are somewhat of a surprise given that the local metamorphic basement yields ages with apparent ages between 1400 and 1650 Ma (Karlstrom et al., 1997). However, numerous samples from several units corroborate this finding and support the interpretation that the Doye Formation is younger than 1140 Ma. Based on the entire data set, and petrographic and microprobe examination, it is not likely these young muscovite grains represent alteration ages or ages that have been thermally affected by, for instance, 1.1 Ga magmatism or volcanism.

Detrital zircon age analyses corroborate the mica results and show that much of the Unkar Group (excluding the Bass Formation) was deposited between 1100 and 1170 Ma. U-Pb geochronology of detrital zircons was conducted by laser ablation multicollector–inductively coupled plasma–mass spectrometry (LA-ICPMS), utilizing a Micromass Isoprobe and a New Wave DUV 193 laser ablation system. The common Pb correction is made by using the measured \(^{206}\)Pb and assuming an initial Pb composition from Stacey and Kramers (1975). Errors that propagate from the measurement of \(^{206}\)Pb, \(^{207}\)Pb, \(^{208}\)Pb, \(^{209}\)Pb, and \(^{204}\)Pb are reported at the 1σ level. Additional errors that affect all ages include uncertainties from (1) U decay constants, (2) the composition of common Pb (assumed to be ±1.0 for \(^{206}\)Pb/\(^{204}\)Pb and ±0.3 for \(^{207}\)Pb/\(^{204}\)Pb), and (3) calibration.
Figure 9. (A–C) Age spectrum and K/Ca diagrams for biotite and hornblende from Grand Canyon diabase sills. (D) Isotope correlation diagram for hornblende sample H98-131.6-4. The hornblende and biotite from H98-131.6-4 yield flat spectra for greater than 90% of the $^{39}$Ar released, but have significantly different apparent ages. Isochron analysis of the hornblende suggests excess Ar contamination for the hornblende, and the preferred emplacement age of the H98-131.6-4 sill is given by the biotite at 1104.3 ± 1.4 Ma. The complex biotite spectrum for 98GC25 is interpreted to result from $^{39}$Ar recoil, and the total gas age of 1113.5 ± 1.4 Ma is tentatively interpreted as the intrusion age. MSWD—mean square of weighted deviates.
Figure 10. (A) Relative probability diagrams of $^{40}$Ar/$^{39}$Ar muscovite age determinations from the Dox Formation. $^{40}$Ar/$^{39}$Ar ages are determined from single crystal detrital muscovites from the Escalante Creek Member of the Dox Formation. Diagram combines 56 total fusion ages with 26 plateau ages from low-resolution step heating of single crystals. The step-heating data reveal that total gas ages are typically 0.8% younger than plateau ages, and therefore all total fusion results have been increased by this percentage. The well-developed node at 1140 Ma indicates that Dox deposition occurred post–1140 Ma, but prior to eruption of Cardenas lavas and intrusion of diabase (vertical line) at ca. 1115–1104 Ma. (B–E) Combined concordia and relative probability plots of detrital zircons from Grand Canyon Unkar Group and Nankoweap Formation. Zircons compliment detrital mica age determinations and also suggest a strong Grenville-age source for sediment, as well as sources from expected 1450–1350, 1750–1650, and 1840 Ma local basement, and minor Archean grains.
correction. These systematic errors add an additional 2% (1σ) uncertainty to $^{206}$Pb/$^{238}$U and $^{206}$Pb/$^{230}$Pb ages.

One-hundred zircon grains were analyzed from four samples from the Hakatai, Shinumo, Dox, and Nankoweap Formations (Fig. 10). Forty-eight analyses with $>$20% discordance or $>$10% reverse discordance have been eliminated from consideration, leaving 352 reliable ages. This subset of detrital zircon analyses yielded minimum ages for samples that were determined from the youngest cluster of age determinations. Sample LC-02-81-2 was collected from the upper Hakatai Formation and yielded eighty-five zircon grains for analysis. The youngest cluster of ages suggests that the maximum age of deposition was ca. 1187 Ma (Fig. 10B). Interestingly, this sample was dominated by zircons of Grenville age (1200–1350 Ma) and appeared to have little input from more proximal Paleoproterozoic and early Mesoproterozoic crustal rocks.

Sample T01-75-4 from the Shinumo Sandstone yielded eighty-nine zircons representative of a more cosmopolitan source terrain. The maximum age of deposition for the Shinumo is ca. 1176 Ma based on these analyses (Fig. 10C). This sample contained zircons similar to the upper Hakatai sample and is interpreted to reflect continued input from Grenville-age crust. In addition, the Shinumo sample had abundant zircons that reflected proximal crustal ages, suggesting a mixed source.

Sample T02-75-1z was collected from the Escalante Creek Member of the Dox Formation. This sample contained ninety-three zircons for analysis and yielded a maximum age of ca. 1165 Ma, similar to the inferred maximum age of the Shinumo (Fig. 10D). This sample also contained numerous zircons of Paleoproterozoic and early Mesoproterozoic (ca. 1.4 Ga and 1.6–1.8 Ga) source rocks, but also contained significant numbers of zircons that are older than proximal crustal sources. This implies that there has been recycling of old zircons from proximal Paleoproterozoic metasedimentary rocks, or zircons have been transported from some distant older crustal block, presumably from the south, based on paleocurrent analysis.

Sample K00-53-3 was collected from the upper member of the Nankoweap Formation and yielded 85 zircons for analysis. The zircons were considerably younger than the underlying Unkar Group and suggested a maximum age of ca. 942 Ma (Fig. 10E). This maximum age is probably conservatively old, and the actual age of the upper Nankoweap Formation may be closer to inferred ages of ca. 850–900 Ma determined from paleomagnetic studies (Lucchitta and Hendricks, 1983; Weil et al., 2003).

The lower Nankoweap Formation (ferruginous member) remains undated, and could potentially be older than 942 Ma, depending on the duration of the hiatus separating lower and upper Nankoweap rocks.

**TIMING OF DEFORMATION IN THE GRAND CANYON SUPERGROUP**

Sears (1973) recognized three main types of faulting in the Unkar Group: (1) faults related to intrusion of ca. 1.1 Ga diabase, (2) faults related to regional shortening, and (3) faults related to extension and domino-style tilting of Unkar strata. Like Elston (1979), Sears interpreted the tilting of Unkar strata on NW-striking faults to coincide with the Late Precambrian “Grand Canyon Revolution” of Maxson (1961). Sears (1973) also described NE-striking, steeply dipping contractual faults that folded Unkar strata into monoclines. These monoclines are clearly truncated by the Middle Cambrian Tapeats Sandstone, indicating that these structures are also Precambrian in age, but the relative importance of the extensional and contractual faults was not well understood.

More recent work in the Chuar and Unkar Groups (Timmons et al., 2001) has further refined our understanding of multiple, but punctuated episodes of deformation and sedimentation and has led us to abandon the concept of a single Neoproterozoic Grand Canyon orogeny. Instead, rocks of the Grand Canyon Supergroup record multiple extensional events separated by nearly 300 m.y. of geologic time (Timmons et al., 2001). The details of the older, 1300–1100 Ma, deformational events recorded in the Unkar Group, as presented in this paper, suggest a regional tectonic response to progressive plate-margin deformation in the late Mesoproterozoic.

**Lower Unkar Group NW-SE Contraction (>1140–1250 Ma)**

Contractional faults that offset Unkar Group deposits are located in side canyons along the Colorado River, including Red, Vishnu, Bright Angel, and Bass Canyons (Fig. 2). Unkar Group rocks in these locations are folded into monoclines that trend NE and face NW (nearly orthogonal to known N-NW–trending Laramide monoclines). Monoclines in Vishnu and Bright Angel Canyons follow the structural grain of the Granite Gorge Metamorphic Suite, whereas in Bass Canyon the monocline crosscuts the metamorphic grain at high angles. All monoclines have a stratigraphic separation less than 200 m and are beveled by upper Unkar or Tapeats strata.

Field observations suggest that contractual faulting of the Unkar Group predates the 1.1 Ga magmatic activity in the area. In Bright Angel Canyon, 1.1 Ga dikes intrude along NE-striking reverse faults and feed sills within the Hakatai Shale, however, neither the sills nor the dikes appear to be offset by Precambrian contractual movement. In Bass Canyon, mapping shows the monocline truncated by a large dike presumed to be part of the 1.1 Ga magmatism. Precambrian reverse faults and monoclines are not observed to affect any deposits younger than the Shinumo Sandstone. In Red Canyon, the monocline dies out in the Hakatai Shale; in Vishnu Canyon, the monocline diminishes within the Hakatai Shale; in Bright Angel Canyon the monocline penetrates the Shinumo Sandstone and is truncated by the Tapeats Sandstone; and in Bass Canyon the well-developed monocline in the Bass Formation and Hakatai Shale is covered by undeformed Shinumo Sandstone (Fig. 2).

Inspection of depositional patterns of the Bass Formation highlights unusual trends in facies and thickness distribution that suggest a complex interplay between deposition and tectonism. The formation thickens to the west with important and abrupt thickness changes in Vishnu and Bright Angel Canyons that suggest syndepositional response to preexisting relief on the basement surface due to faulting (Fig. 4). The presence of conglomerate and breccia intimately interbedded with carbonate rocks of the Bass Formation suggests local relief, recycling of older carbonate beds, and possibly erosion of basement rocks during deposition. Clast size and composition of conglomerate beds in the Bass Formation suggest high-energy transport of local and distal basement clasts prior to deposition of carbonate beds and recycling of lower carbonate and conglomerate beds during deposition of upper conglomerate lenses. Field observations suggest that faulting plays some role in the development of relief between different crustal blocks that influenced depositional patterns and facies associations.

Faulting during Bass Formation deposition is documented by detailed mapping and measured sections. In Red Canyon, a gentle monocline (30° dip in ramp) folds rocks of the Bass Formation and Hakatai Shale. At this location, the fault does not penetrate exposed beds of the Bass Formation, however, the trace of the monocline trends toward the northeast. Siliciclastic beds of the upper Bass Formation preserve very gentle (10°) angular pinch-outs in the ramp of the monocline (Fig. 11). Transection surfaces are discrete and no lag deposits were observed in the section. The pinch-outs are consistent with erosion of the hanging-wall block during progressive monocline
development and burial of the disconformity by subsequent sedimentation.

In Vishnu Canyon, deposits of the Bass Formation are observed to thin dramatically onto the hanging wall of the monocline (Fig. 12). The reverse fault clearly offsets the lowermost deposits of the Bass Formation (including Hotauta Member) with no obvious changes in thickness. This is, however, not the case in the upper part of the section. On the west, the section is thick, with multiple beds of conglomerate and sandstone. Correlative beds on the east side of the fault lack similar thickness of conglomerate and sandstone; in fact, upper conglomerate beds thin to zero thickness, and sandstone beds thin to a few centimeters thick (Fig. 4). Furthermore, we observed soft sediment slump features and olistostrome deposits in associated carbonate and interbedded siliciclastic rocks over this same horizon (Fig. 12B–C). Sedimentary horizons in the Bass and Hakatai above the attenuated section do not show similar thickness changes, suggesting a hiatus in fault activity. Beds above the disturbed zone are, however, also folded by the monocline, implying renewed Proterozoic movement that postdated these upper deposits. The amplitude of the monocline diminishes into the Hakatai Shale, and the Tapeats Sandstone ultimately truncates the structure. Therefore movement across this structure took place during lower Bass deposition and again sometime between deposition of the Hakatai and Tapeats.

In the Bass Canyon monocline (Fig. 2), folding of bedding took place before Shinumo Sandstone deposition. The Bass Formation is tightly folded into a NW-facing monocline, with beds that are vertical to overturned. Stratigraphic units within the Bass Formation are correlated across the structure without changes in facies distribution or thickness (Fig. 4), suggesting movement after deposition of the Bass Formation. Farther upsection in the Hakatai Shale to Shinumo contact, we observe a well-developed monocline with beds that dip as steep as 45° to the NW. Overlying Shinumo beds are in fault contact with this monocline but show no evidence of folding, suggesting this monocline dies out in the upper Hakatai. The monocline also is truncated by a NW-trending graben that juxtaposes undeformed diabase with overturned beds of the Bass Formation, further showing that the monocline predates 1100 Ma magmatism. Collectively these observations suggest that this monocline developed after deposition of the Bass Formation and before final deposition of the Shinumo Sandstone.

The duration of contractional deformation remains difficult to determine. The apparent conformity of Dox with Shinumo deposition suggests that contraction was long-lived and lasted from at least 1250 Ma until ca. 1165 Ma (assuming little difference in age between Shinumo and Dox deposits). Huntoon and Sears (1975), report 183 m of east-side-up reverse-sense movement along the Bright Angel fault zone after emplacement of ca. 1100 Ma diabase sills and dikes. If accurate, this would imply that contractional deformation was concurrent with diabase intrusion and orthogonal extensional deformation.

**Uncar Group NE-SW Extension ca. 1250–1150 Ma**

The main extensional deformation that tilted Grand Canyon Supergroup strata before deposition of the Cambrian-age Tapeats Formation is well known (Powell, 1875; Walcott, 1889). Uncar Group rocks are cut by Proterozoic normal faults of variable displacement that strike NW and commonly form full or half grabens (Fig. 2). Uncar Group strata generally dip northeast toward steeply southwest-dipping normal faults and form coherent ≤20°-dipping tilt blocks with half-graben geometries.

The timing of this deformation has been variably interpreted as late Neoproterozoic (Elston, 1979) and late Mesoproterozoic (Sears, 1973). Our data are interpreted to suggest that early extensional deformation overlapped with, but was generally younger than, contractional deformation, and that the main tilting and normal faulting of the Unkar Group took place before deposition of the ca. 900 Ma upper Nankoweap Formation and Chuar Group (Fig. 3). Mutually crosscutting relationships between normal faults and diabase sills and dikes (Sears, 1973) suggest that mafic magmatism at ca. 1.1 Ga overlapped in time with extensional faulting in the Unkar Group (Fig. 13A). Intraformational faults near the Palisades fault die out in the Cardenas Basalt (Fig. 13B). There is a 3°–5° angular unconformity between the Unkar Group and Nankoweap Formation (Gebel, 1978; Fig. 13C), and in the Tanner graben several Unkar-age normal faults die out upsection in the Nankoweap Formation (Gebel, 1978; Fig. 13D). Some of these observations are discussed in more detail in the following.

The Palisades fault is a key fault for deciphering pre-Chuar Group extension (Fig. 2). The fault strikes 310° and dips steeply to the southwest. Proterozoic stratigraphic separation across this structure is ~1100 m down-to-the-southwest after ~300 m of Laramide reverse slip is restored. Timmons et al. (2001) argued that the continuity of Butte fault hanging-wall strata and discontinuity of Butte fault footwall strata across the Palisades fault indicate that the Palisades fault is truncated by the Butte fault and that faulting and tilting of Unkar Group rocks predate Butte fault movement and Chuar Group deposition. Corollary observations of intraformational faults subordinate to the

![Figure 11. Field photo of the ramp of a Precambrian monocline in Red Canyon. View is toward the NE, and beds of the uppermost Bass Formation dip gently to the NW. Sedimentary pinch-outs of beds are observed indicating monoclinal development, and erosion of mildly deformed depositional surfaces was contemporaneous with deposition of the Bass Formation. The observed thinning of beds and sedimentary pinch-outs are interpreted to reflect little or no structural thinning. All sedimentary facies show a thinning onto the monocline, and the gentle dips suggest little structural modification. This suggests that NW-directed shortening and deposition were synchronous.](image-url)
Palisades structure (Fig. 13B) and Basalt Canyon fault support the interpretation that faulting, igneous activity, and sedimentation were intimately linked. Post–Unkar Group movement of the Palisades and other large extensional faults occurred to further tilt and ultimately structurally isolate Unkar outcrops. Interestingly, the Neoproterozoic-age Chuar Group does not share this same history of tilting and graben formation. Rather, Chuar rocks record the development of a N-S–trending Neoproterozoic growth syncline (Timmons et al., 2001).

The angular discordance between the Nankoweap Formation and the Unkar Group indicates a period of faulting and rock tilting after eruptions of the Cardenas Basalt and prior to deposition of Nankoweap strata (Fig. 13C). Here, the 3°–5° angular discordance between the two rock packages accounts for part of the total 10°–15° tilt preserved by Unkar deposits. Extension recorded within the undated lower Nankoweap Formation seems to be a continuation of Unkar-related extension, as supported by the similarity between red-beds in the Dox Formation, interflow red-beds in the Cardenas Basalt, and red-beds of the lower Nankoweap Formation (Elston, 1979).

Evidence for extensional deformation is recorded by unconformities and intraformational faults in the lower Nankoweap Formation. Elston and Scott (1976) and Link et al. (1993) reported a major unconformity within the Nankoweap Formation and suggested that faulting and erosion preceded deposition of the upper member of the Nankoweap Formation. Intraformational normal faults within the Tanner graben are truncated by strata of the upper Nankoweap Formation and Chuar Group (Fig. 13D), suggesting extension during early phases of Nankoweap deposition. Adjacent to major faults, such as the one in Basalt Canyon, sedimentary beds pinch out against the fault. Further, a coarse-grained (5–10 mm clasts) lag deposit (<1 cm thick) in the Nankoweap Formation in Basalt Canyon contains clasts of dolomite sandstone olistostrome.
TECTONIC INFERENCES FROM THE CA. 1255–1100 MA UNKAR GROUP

the underlying Cardenas Basalt and indicates exposure of the basalt (by faulting) during early Nankoweap deposition. The duration and tectonic importance of the disconformities below, within, and at the top of the Nankoweap Formation remain poorly understood. We suggest, however, that the lower Nankoweap Formation records a continuation of Dox-age deposition and faulting that predates ca. 950 Ma deposition of the upper Nankoweap Formation.

The onset of extensional deformation in the Unkar Group has been difficult to determine, but may have begun early during Unkar deposition. The Phantom graben (Fig. 2) is a NW-striking symmetrical graben that juxtaposes rocks of the lower Unkar Group with basement metamorphic rocks. The Bass Formation at this location exhibits unusual thickness and facies changes suggestive of synsedimentary faulting. Figure 14 shows the Bass Formation through the lower Hakatai Shale bounded by faults that strike to the northwest. At the base of the section, competent beds of dolomite are offset by intraformational normal faults. Fault offsets are estimated between 10 and 20 m of stratigraphic separation within the Bass Formation. Siliciclastic beds of the upper Bass Formation are observed to be laterally continuous across the entire width of the graben, suggesting a truncation of subordinate fault sets. This suggests that contractional and extensional deformation were contemporaneous and record a single strain field during Bass deposition.

**Fault Kinematics**

Orientations of normal faults in the Grand Canyon Supergroup are shown in Figure 15. Normal faults in the Unkar Group and Chuar Group occur in two main populations that have northwest to north strikes and are steeply dipping. These faults are Precambrian in age, because large offset normal faults can be traced to the Tapeats or older sedimentary unit contact where they are truncated, and in Phanerozoic
rocks, faults of similar orientation preserve reverse sense of movement. Both Unkar and Chuar Groups show conjugate faults (Fig. 15). Paleostress axes for Unkar fault planes record subvertical $\sigma_1$, subhorizontal $\sigma_2$ that trends toward 306°, and subhorizontal $\sigma_3$ that has a bearing of 037° (Fig. 15). Chuar Group faults also record subvertical $\sigma_1$; however, the orientation of $\sigma_2$ is 334° and $\sigma_3$ is 065° (Fig. 15). The difference in fault plane populations is interpreted as overprinting of Meso- and Neo-proterozoic fault populations. Extension during Chuar deposition had shifted to E-W extension.

**REGIONAL SEDIMENTARY AND TECTONIC CORRELATIONS**

Our updated geochronology allows us to confirm and refine proposed regional correlations of Upper Mesoproterozoic sedimentary successions and extensional events. Earlier attempts to correlate regionally dispersed sedimentary successions relied heavily on lithologic correlations and paleomagnetic evaluations of red-bed deposits and igneous rocks. Link et al. (1993) reported that onset of Unkar deposition occurred at ca. 1250 Ma, based on an interpolation of Bass Formation paleomagnetic pole in the 1300–1200 Ma North American apparent polar wander (APW) path, which is in agreement with our new geochronology. They also noted that paleomagnetic pole positions from the upper Unkar Group define a counterclockwise, NNE-elongated loop that resembles the loop defined by Keweenawan rocks of the mid-continent rift system (Halls, 1974; Pesonen and Halls, 1979).

**Previous Late Mesoproterozoic Regional Correlations**

Lithologic correlations between the Apache Group and Unkar Group have long been postulated. Shride (1967) suggested that the Mescal Limestone (including carbonate and overlying argillite member) and the Bass Formation and Hakatai Shale were regional correlatives. He also recognized similarities between units of the Shinumo Sandstone and Troy Quartzite, including “curiously twisted and gnarled” beds observed by Powell (1875) and Noble (1914, p. 51) in the Grand Canyon similar to the Che-diski Member of the Troy. No lithologic correlations have been postulated between the Pioneer Shale and Dripping Spring Formations in the Apache Group and Grand Canyon rocks. Likewise, it has been suggested that the Dox has no depositional equivalents in the Apache Group (Shride, 1967; Wrucke, 1989).

Further regional correlations between other late Mesoproterozoic successions have been attempted. Wrucke (1966) suggested that carbonate rocks of the Middle Crystal Spring Formation of the Pahrump Group correlate to the Bass and Mescal Formations in Arizona, based on lithologic similarities and similar diabase intrusive relationships. The correlations based on the combined data set (paleomagnetic and lithologic correlation) are supported by our geochronology from the Unkar Group.

**Proposed Regional Correlation and Tectonic Model**

Our new geochronology indicates that sedimentary rocks over a large region of the southwestern United States were deposited synchronously and share a common sedimentologic and tectonic history. We postulate that the depositional record of the Pioneer Shale and Dripping Spring Quartzite mostly predates the sedimentary record in the Grand Canyon Unkar Group, and may record part of the exhumation history of the Grand Canyon region (Karlstrom et al., 1997). Rocks of the Pioneer and Dripping Spring Formations record derivation of coarse clastics from some unknown northern basement source (Cullom, 1996). Best estimates for the age of the Pioneer Formation come from detrital zircon studies and suggest that this section is ca. 1328 ± 5 Ma, based on euhedral and clear zircons from a tuff or possibly a reworked tuff (Stewart et al., 2001).

Rocks of the Allamoore Formation, dated at 1250 ±20/–27 Ma, 1253 ± 15 Ma, and 1256 ± 5 Ma, Castner Marble, dated at 1260 ± 20 Ma (Bickford et al., 2000), and Bass Formation, dated at 1254.8 ± 1.6 Ma, appear to be age equivalents, contain numerous dolomitic/marble stromatolite-bearing beds, and have been interpreted to record shallow marine deposition (Figs. 1 and 16; Dalton, 1972; McConnell, 1975; Nyberg and Schopf, 1981; Pittenger et al., 1994). We support the original correlation of Wrucke (1966, 1989) of the middle Crystal Spring Formation carbonate with other ca. 1250 Ma carbonate sequences based on crosscutting 1.08 Ga diabase (Heaman and Grotzinger, 1992), and the recognition of an unconformity between the middle and upper Crystal Spring Formation (Prave, 1998, personal commun.). We also support, after Shride (1967), the correlation of late Mesoproterozoic sedimentary rocks in central Arizona (Mescal Formation), and carbonate facies in subcrops in New Mexico (Debaca Group; Amarante, 2001).

**Figure 14.** In Bright Angel Canyon near Phantom Creek, a NW-striking graben preserves sedimentary rocks of the Bass Formation through lower Shinumo Sandstone. Minor faults within the graben offset lower beds of the Bass Formation and are truncated by siliciclastic beds of the uppermost Bass Formation and lower Hakatai Shale. Fault offsets at the base of the section are estimated at ~10–20 m. This suggests that early faulting of the Bass included movement of NW-striking faults in response to NE-SW extension.
with the ca. 1250 Ma carbonate sequences in the southwestern United States.

Current paleogeographic/depositional models suggest that the pre-orogenic marine rocks in west Texas were deposited during island-arc convergence in a back-arc, orogen-parallel, extensional setting (Rudnick, 1983; Pittenger et al., 1994; Mosher, 1998). Assuming that the 1255 Ma age of the Bass Formation is accurate, we argue that the regional distribution of ca. 1250 Ma carbonates, with mixed siliciclastic beds (with shallow water indicators) and associated stromatolite beds, suggests deposition in shallow marine intertidal to subtidal conditions in a back-arc–epicontinental seaway (Figs. 1 and 16). The record of regional contraction at 1250 Ma in the Grand Canyon, however, may differ from the orogen-parallel extensional back-arc setting for west Texas carbonate successions. Monoclines of the lower Unkar Group were developing during regional deposition of carbonate sequences, and there is an apparent paucity of convergent deformation in west Texas deposits at ca. 1250 Ma. As monoclinal development waned in the Grand Canyon region, contractional deformation, as a large-scale thrust belt during continent-continent phase convergence, is recorded in younger deposits of the Hazel Formation (Bickford et al., 2000).

New detrital mica and zircon ages provide a much better constraint on the timing of deposition of the Dox Formation than previous studies and allow correlation to other sequences in the southwest United States. The 1140–1104 Ma Dox Formation records the transition from shallow marine deposits of quartz arenite of the Shinumo Sandstone to fluvial-deltaic arkosic sandstones of the basal Dox Formation. The transition is very abrupt (over 10 m), but illustrates a regression of the Shinumo sea and burial by fluvial/deltaic facies of the lower Dox. Sedimentary structures in the upper Shinumo Sandstone exhibit general bimodal paleocurrent directions. The basal member of the Dox Formation is characterized by large-scale (tens of meters wide) channel facies sandstone interbedded with intrafluvial mudstone. The channel facies record a consistent paleocurrent direction (determined from trough bedding axes) toward the north that persisted into upper beds in the Dox (Fig. 4). We interpret the change in sandstone composition and change in depositional systems to reflect a fundamental reorganization of the paleogeography related to progressive Grenville contraction.

Late Mesoproterozoic (1.2–1.1 Ga)–age micas in basement rocks of the Southwest United States are rare, as much of the Southwest records cooling histories that predate deposition of the Dox Formation. For instance, basement rocks in the Southwest record cooling through the mica closure temperature window by or soon after 1.4 Ga (Shaw et al., 1999). The few exceptions include the Uncompahgre Uplift (ca. 1100 Ma; Timmons et al., 2002), small portions of central Arizona, and the core of the Grenville orogeny (Bickford et al., 2000). Of the known 1100 Ma mica thermochronologic terranes, the Grenville Front provides the most likely southern source for fluvial detritus of a single provenance during Dox deposition. We conclude that the proximal facies of the Hazel Formation (Soegaard and Callahan, 1994; Bickford et al., 2000) in west Texas and the distal fluvial/deltaic facies of the Dox Formation represent age-equivalent deposits (Fig. 16). This implies also, that during Dox deposition, there was no highland separating Unkar and Apache Group basins. The Dox Formation records intracratonic deformation and transition of regional strain patterns from compression near the orogen to orthogonal extension in the continental interior. The relative timing of contraction and extension in the Unkar Group suggests early NW-directed shortening and coeval perpendicular extension followed by dominant NE-SW–directed extension during Unkar deposition. Similar relationships have been postulated closer to the orogen in the Central Basin Platform (Adams and Keller, 1996), where a preserved Phanerozoic-inverted Precambrian graben is present in subcrop.

Several modern examples of collisional-perpendicular extensional basins exist that share many similar features with this ancient orogenic belt. Mechanisms driving intracratonic deformation remain a topic of debate; however, many workers have concluded that intracratonic rifting was a response to plate-margin convergence. The NE-trending Baikal rift has been interpreted as a passive rift formed in response to the Indo-Eurasian collision (Tapponnier and Molnar, 1979; Hutchinson et al., 1992). Another example includes the upper and lower Rhine grabens in Western Europe. Following Şengör et al. (1978), many workers have refined the timing and kinematics of the Rhine graben, and they have concluded that graben formation
and Alpine tectonics were pencontemporaneous and kinematically linked (Larroque and Laurent, 1988; Laubscher, 2001).

Temporally associated with Grenville orogenesis and regional deformation are widely dispersed mafic intrusive and extrusive igneous rocks (Fig. 1, inset). Mafic igneous rocks span a broad interval in time, from ca. 1235 Ma in the Grenville of Ontario (Bethune, 1997), 1109–1087 Ma in the Midcontinent region (Van Schmus, 1992), to 1165–1120 Ma in west Texas (Keller et al., 1989; Barnes et al., 1999), and new data indicate 1115–1104 Ma in the Grand Canyon. Mafic intrusions commonly occur as sills in both Mesoproterozoic sedimentary rocks and the crystalline basement in the southwestern United States, which led Howard (1991) to postulate that sills record subvertical $\sigma_3$ during NW-directed shortening. NW-striking mafic dikes of Grenville age are also observed over much of Laurentia and are commonly associated with NW-striking rift zones and record NE-SW-directed extension during orogenesis (Keller et al., 1989; Bethune, 1997). In the Grand Canyon, both voluminous subhorizontal mafic sills and NW-striking mafic dikes are observed, suggesting a $\sigma_3$ that is either subhorizontal and NE-trending or subvertical. We suggest that both $\sigma_2$ and $\sigma_3$ were subequal, with perhaps more local stresses controlling the orientation of $\sigma_2$ and $\sigma_3$. Both structural relationships are compatible with NW-directed shortening.

The extent of Grenville deformation in the continental interior becomes clearer when considering seemingly disparate data sets. Laramide contraction in the southwestern United States reactivated a linked network that involved segments of late Mesoproterozoic and Neoproterozoic normal faults in the Grand Canyon (Huntoon, 1971; Sears, 1973; Huntoon et al., 1996; Timmons et al., 2001). Marshak et al. (2000) and Timmons et al. (2001) argued that the strong northwest trend of faults observed over much of Laurentia (Grand Canyon, Central Basin Platform, and Sudbury dikes; cf. Fahrig and West, 1986) records regional northeast extension, which overlaps in time with Grenville-age NW-directed contraction.

Further tests of ancestry of regional fault networks in the Colorado Plateau and Rocky Mountain region are under way via $^{40}$Ar/$^{39}$Ar studies. K-feldspar data document important cooling events that may represent periods of exhumation during (1) Grenville orogenesis (1200–1100 Ma), (2) Late Precambrian rifting at 800–700 Ma (western Cordillera) and perhaps 600–550 Ma (Oklahoma Aulacogen trend), and (3) Ancestral Rockies deformation (ca. 350 Ma). At more local scales, feldspars seem to resolve disparate cooling paths across
Tectonic Inferences from the Ca. 1255–1100 Ma Unkar Group

 discrete faults, suggesting that faults controlled exhumation of different crustal blocks at different times (Timmons et al., 2002; Timmons, 2004). The occurrence of a 1250–1100 Ma cooling trend over much of the Rocky Mountain region and Grand Canyon suggests a dynamic response in the continental interior to Grenville orogenesis.

SUMMARY

The lower Unkar Group records contemporaneous contractional and extensional deformation and deposition. NE-trending monoclines deformed lower units of the Unkar Group and developed during sedimentation. Mapping suggests that convergent structures deformed mostly rocks of the lower Unkar Group and not the Neoproterozoic-age Chuar Group. Hence, combined with sedimentologic evidence, contractional structures are interpreted to be late Mesoproterozoic in age. The Unkar Group is also faulted and tilted by NW-striking normal faults that record NE-SW-directed extension prior to Chuar Group deposition. NE-directed extension was the dominant kinematic regime during late Unkar and early Nankoweap deposition. Unkar Group deformation is temporally and kinematically linked to Grenville orogenesis. In contrast, extensional deformation recorded in the ca. 800–740 Ma Chuar Group records E-W-directed extension related to the incipient rifting of the Western Cordillera of Laurentia, nearly 200 m.y. after Unkar deformation and tilting (Timmons et al., 2001). The terms Grand Canyon “Revolution” of Maxson (1961) and “Orogeny” of Elston (1979) refer to a single extensional event, but there were at least two separate deformational episodes linked to supercontinent assembly (ca. 1250–950 Ma) and rifting (ca. 800–750 Ma).

New geochronology confirms and refines published correlations for late Mesoproterozoic rocks in the Southwest. Carbonate rocks were deposited in an epicontinental seaward unconformity, Laramide-style convergence in the continental interior at ca. 1250 Ma. Subcrop rocks of the Las Animas Formation, Central Basin Platform, and DeBaca terrace correlate to exposed sections that include the Allamore, Tumbledown, and Hazel Formations of west Texas, the Apache Group of central Arizona, the lower and middle Crystal Springs Formation of Death Valley, California, and the Unkar Group of the Grand Canyon. Data from the Grand Canyon indicate pulses of sedimentation with discrete episodes of tectonism associated with the evolving Grenville orogen from ca. 1250–1100 Ma to perhaps as young as 950 Ma, which probably also affected correlative rocks of the region.

ACKNOWLEDGMENTS

We would like to acknowledge the following collaborators who have been part of the overall scientific development of the Grand Canyon Supergroup project: Carol Dehler, John Bloch, Tony Prave, Steve Cather, Adam Read, Stacy Wagner, Paul Bauer, Mary Simons, Colin Shaw, Brad Igl, Mike Doe, Jake Armour, Casey Cook, Sarah Tindall, Arlo Weil, John Geissman, Annie McCoy, Micah Jessup, Mark Quigley, Andy Stone, Lisa Peters, Tim Lite, Andy Knoll, Suzanna Porter, and Jason Rauci. This work was made possible by the National Science Foundation grant EAR-9706541 to Karl Karlstrom, John Geissman, and Maya Elrick, and EAR-9902955 to Karlstrom and Heizler for thermochronology efforts. Further support comes from Ben Donegan (consulting geologist), Conoco, Inc., and Schlumberger, Inc. We would also like to acknowledge the helpful reviews of Gerry Ross and Jim Sears.

REFERENCES CITED


Reed, V.S., 1976, Stratigraphy and depositional environment of the upper Precambrian Hakkatai Shale, Grand Canyon region [M.S. thesis]: Flagstaff, Northern Arizona University, 163 p.


