Detrital zircon U-Pb geochronology of Paleozoic strata in the Grand Canyon, Arizona

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ABSTRACT

We determined U-Pb ages for detrital zircons from 26 samples of Paleozoic sandstone from the Grand Canyon. Cambrian strata yield mainly ca. 1.44 and 1.7–1.8 Ga ages that indicate derivation from nearby basement rocks of the Yavapai Province. Devonian strata contain zircons of 1.6–1.8 Ga, 1.34–1.40 Ga, and ca. 520 Ma, suggesting derivation from the Mazatzal and Yavapai Provinces, midcontinent region, and the Amarillo-Wichita uplift, respectively. Mississippian strata record a major change in provenance, with predominantly 415–475 Ma and 1030–1190 Ma grains interpreted to have been shed from the central Appalachian orogen. Pennsylvanian strata contain subequal proportions of 1.4–1.8 Ga grains derived from basement rocks exposed in the Ancestral Rocky Mountains and 409–464 and ca. 1070 Ma grains derived from the Appalachians. Permian strata contain abundant Appalachian zircons, including 270–380 Ma grains, and a lesser proportion of grains derived from the Ancestral Rocky Mountains. Transcontinental transport during Mississippian through Permian time is interpreted to have occurred in large river systems, facilitated by northeasterly trade winds during low sea level and by coastal currents. A compilation of young ages from all Upper Paleozoic strata yields age peaks of 270–365 Ma, 395–475 Ma, and 515–640 Ma, an excellent match for Alleghanian, Acadian, Taconic, and Neoproterozoic (peri-Gondwanan) episodes of magmatism along the Appalachian margin. Lag times of the youngest grains in these Upper Paleozoic strata average ~25 m.y., suggesting relatively rapid exhumation and erosion of Appalachian source regions.

INTRODUCTION

The spectacular exposures of Paleozoic strata in the Grand Canyon (Fig. 1) provide an excellent opportunity to determine the provenance of Cambrian through Permian sandstones of the southwestern United States. These strata have been studied by many geologists during the past 140 yr, including the pioneering work of J.W. Powell, G.K. Gilbert, C.E. Dutton, C.D. Walcott, and E.D. McKee, which has set the stage for many more detailed studies that are reviewed in a comprehensive fashion by Beus and Morales (2003).

Although there have been many thorough and detailed analyses of Grand Canyon strata, there are few constraints on the provenance of sandstones that make up much of the ~1300 m of Paleozoic stratigraphy (Fig. 2). For Cambrian strata, regional studies of paleogeography, facies patterns, and paleotransport indicators (e.g., cross-beds) have suggested that most of the sand was derived from Precambrian basement within and east of the Grand Canyon (e.g., McKee and Resser, 1945; and as summarized by Middleton et al., 2003). Similar criteria from Pennsylvanian and Permian sandstones generally supported the conclusion that most sand was derived from the Ancestral Rocky Mountains of Colorado, New Mexico, Utah, and Arizona and/or from basement uplifts farther north (Peterson, 1988; Blakey, 1988; Blakey et al., 1988; Marzolf, 1988; Johansen, 1988). Johansen (1988) considered the Appalachian orogen as a possible source but preferred derivation from northern cratonic sources based on paleogeographic considerations. In contrast, Dickinson and Gehrels (2003) presented detrital zircon analyses from the Coconino Sandstone and concluded that the Appalachian orogen was a likely source for this unit. This conclusion has been supported by regional stratigraphic analyses presented by Blakey (2008, 2009a).

This study uses detrital zircon geochronology to place additional constraints on the origin and transport history of Cambrian, Devonian, Mississippian, Pennsylvanian, and Permian sandstones in the Grand Canyon. Our approach involves determining the ages of zircon crystals from each unit, comparing these age distributions with the ages of rocks in possible source terranes, and using available facies, paleocurrent, and paleogeographic information to reconstruct possible pathways of sediment transport. The results of these analyses in some cases confirm the conclusions of previous workers, and in other cases provide new provenance interpretations.

METHODS

Sandstone samples ranging from 2 to 10 kg in weight were collected from ~50-cm-thick horizons that are representative in terms of grain size and composition. The samples were then processed and analyzed using procedures outlined by Gehrels (2000, 2011) and Gehrels et al. (2006, 2008), which are designed to produce a final age distribution that accurately reflects the true distribution of detrital zircon ages in each sample. Initial steps included using a jaw crushe by guest

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procedures (mainly slower processing across Wilfl ey table) were used, and zircons down to ~20 µm in diameter were retained and analyzed. Impurities in the final mineral separates were removed by Wig-L-Bug shaking to break apart softer grains and by handpicking. A representative split of the final zircon yield was mounted in a 1″ (2.54 cm) diameter epoxy plug along with crystals of our SL (Sri Lanka) zircon standard (563.5 ± 3.2 Ma; Gehrels et al., 2008).

Prior to analysis, cathodoluminescence (CL) images were generated for most samples in an effort to locate laser pits in homogeneous portions of crystals. Where multiple age zones were identified, ablation pits were located in the oldest domain of sufficient size for analysis. For several samples that were analyzed without the benefit of CL images, images were acquired after acquisition, and the isotopic analyses were evaluated in light of this spatial information.

Figure 1. (A) Location of samples in the Grand Canyon (shaded area). (B) Exposure of ~1310 m of Paleozoic strata along the north rim of the central Grand Canyon, with formation names indicated.

Fortunately, the number of analyses conducted on multiple domains was small, and all analyses conducted on multiple domains were rejected using the criteria outlined below.

Approximately 100 analyses were conducted from each sample. Grains were selected for analysis at random, and crystals were omitted only if they were too small for analysis with this beam diameter, or if they did not contain a portion that was free of fractures or inclusions.

Analytical complexities were evaluated for all samples during data acquisition by comparison of the time-resolved $^{206}$Pb/$^{238}$U of standards versus unknowns, and examples of different types of complexities are shown on Figure 3. Given that the standard zircons are homogeneous, their pattern of changing $^{206}$Pb/$^{238}$U
values during an analysis was used to define the pattern of fractionation with increasing laser pit depth. Unknown zircon crystals that are homogeneous (or are analyzed from a homogeneous domain) generate a similar pattern of $^{206}\text{Pb}/^{238}\text{U}$ fractionation (e.g., Figs. 3A and 3B). In contrast, analysis of multiple domains from complex crystal yields a $^{206}\text{Pb}/^{238}\text{U}$ pattern that is commonly different from the standard zircon. Approximately 10% of the analyses conducted displayed such complexities (e.g., Pb loss, inheritance, analysis of multiple domains) will tend to scatter analyses from their true age. The ranges and peak ages of the main clusters in each sample were determined with an in-house routine that determines (1) the age ranges containing continuous age probability from at least three analyses (uncertainty represented at the 2σ level), and (2) the ages of all peaks in age-probability domain that include at least three constituent analyses. The output from this routine is included for each sample, and for each set of samples, in Table DR1 (see footnote 1).

The ages were also filtered during interpretation in that the greatest significance is placed on analyses that belong to clusters (e.g., Gehrels, 2011). This strategy tends to reduce the impact of compromised analyses, because all sources of complexity (e.g., Pb loss, inheritance, analysis of multiple domains) will tend to scatter analyses from their true age. The ranges and peak ages of the main clusters in each sample were determined with an in-house routine that determines (1) the age ranges containing continuous age probability from at least three analyses (uncertainty represented at the 2σ level), and (2) the ages of all peaks in age-probability domain that include at least three constituent analyses. The output from this routine is included for each sample, and for each set of samples, in Table DR1 (see footnote 1).

The uncertainties shown on Table DR1 (see footnote 1) and all accompanying figures include only the internal (random or measurement) errors. External (systematic) errors are reported in Table DR1 (see footnote 1) for each sample, and these include contributions from uncertainties in standard analyses, age of the standard, composition of common Pb, and U decay constants. For these samples, the external uncertainties average 1.4% for $^{206}\text{Pb}/^{238}\text{U}$ ages and 1.1% for $^{206}\text{Pb}/^{235}\text{U}$ ages (standard deviations at 2σ level).

All analyses are shown graphically on Pb*/U concordia plots (Fig. 4) and on both relative probability density plots (age-distribution diagrams) and cumulative probability density plots (Fig. 5). The relative probability density plots were constructed by simply summing all ages and their uncertainties and normalizing plots such that all curves contain the same area.

Age distributions were also compared using the K-S statistic (Press et al., 1986), which compares the observed difference between two age distributions against the difference predicted from the number of analyses. The measure of difference is the $P$ value, which is 1.0 if the age distributions are identical and 0.0 if there are few elements in common. The critical $P$ value for distinguishing two age distributions at the 95% confidence level is 0.05—higher $P$ values indicate that the two age distributions are significantly different, and lower $P$ values indicate decreasing similarity. For these comparisons, analytical uncertainties of each analysis are included, which results in higher $P$ values and therefore a more conservative comparison.

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**Figure 2. Schematic column of the Paleozoic strata present in the central Grand Canyon (adapted from R. Blakey, http://jan.ucc.nau.edu/~rcb7/Kaibab_Trail.jpg). Sample names are indicated.**

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The essential isotopic information and ages for acceptable analyses are reported in Table DR1 (see GSA Data Repository1). Analyses that are not included are those rejected during data acquisition or reduction (as described already) or analyses with high error (>10% uncertainty in either $^{206}\text{Pb}/^{238}\text{U}$ age or older than 1.0 Ga $^{206}\text{Pb}/^{207}\text{Pb}$ age) or excessive discordance (>20% discordant or >5% reverse discordant). These rather generous cutoffs were selected in an effort not to bias the final set of ages—a more rigid uncertainty cutoff would tend to eliminate younger analyses due to low Pb signal intensities, whereas a tighter discordance cutoff would tend to eliminate older analyses due to enhanced probability/degree of Pb loss with increasing age (Gehrels, 2011).

The ages were also filtered during interpretation in that the greatest significance is placed on analyses that belong to clusters (e.g., Gehrels, 2011). This strategy tends to reduce the impact of compromised analyses, because all sources of complexity (e.g., Pb loss, inheritance, analysis of multiple domains) will tend to scatter analyses from their true age. The ranges and peak ages of the main clusters in each sample were determined with an in-house routine that determines (1) the age ranges containing continuous age probability from at least three analyses (uncertainty represented at the 2σ level), and (2) the ages of all peaks in age-probability domain that include at least three constituent analyses. The output from this routine is included for each sample, and for each set of samples, in Table DR1 (see footnote 1).

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Figure 3. Cathodoluminescence (CL) images showing measured $^{206}\text{Pb}^{\ast}/^{238}\text{U}$ as a function of types of complexities in zircon grains. Upper images are color CL images that have been adjusted only for brightness and contrast. Scale bars in lower left are 20 µm in length. Lower diagrams show changes in $^{206}\text{Pb}^{\ast}/^{238}\text{U}$ during the 15 s of data acquisition, with lower value of zero and variable upper values. (A) Ablation pit and measured $^{206}\text{Pb}^{\ast}/^{238}\text{U}$ for a grain with a single CL domain. (B) Zircon grain with multiple domains. Measured $^{206}\text{Pb}^{\ast}/^{238}\text{U}$ shows that entire analysis was conducted within a single domain. (C) Zircon grain with ablation pit overlapping two domains. $^{206}\text{Pb}^{\ast}/^{238}\text{U}$ shows an increasing proportion of older zircon with increasing pit depth. (D) Zircon crystal that appears homogeneous in CL, whereas change in $^{206}\text{Pb}^{\ast}/^{238}\text{U}$ shows that a younger domain was intersected at depth. (E) Zircon grain that appears homogeneous in CL, whereas a dip in $^{206}\text{Pb}^{\ast}/^{238}\text{U}$ likely represents the intersection of a fracture or inclusion. Zircon grains and analyses are from Surprise Canyon Formation sample 3.

Figure 4. $^{207}\text{Pb}^{\ast}/\text{U}$ concordia diagram of ages (Ma) of zircon grains from each sample. Uncertainties are shown at the 1σ level. Diagram was constructed with the use of Isoplot (Ludwig, 2008).
All concordia diagrams were generated with Isoplot (Ludwig, 2008), whereas the normalized and cumulative age-probability plots, peak/range values, and K-S analyses were generated with routines available from the Arizona LaserChron Center Web site (http://www.laserchron.org).

PALEOZOIC STRATA OF THE GRAND CANYON

The following descriptions are based primarily on information summarized in Beus and Morales (2003) and are keyed to the stratigraphic column shown in Figure 2 (adapted from R. Blakey, http://jan.ucc.nau.edu/~rcb7/Kaibab_Trail.jpg).

Tapeats Sandstone (Lower and Middle Cambrian)

The Tapeats Sandstone is the lower unit of the Tonto Group, a classic transgressive sequence of Cambrian age (Fig. 2). These strata rest unconformably on Precambrian crystalline basement of the 1.75–1.73 Ga Granite Gorge metamorphic suite (Karlström et al., 2003), various Paleoproterozoic and Mesoproterozoic plutons, and Middle–Upper Proterozoic strata of the Grand Canyon Supergroup (Hendricks and Stevenson, 2003; Ford and Dehler, 2003). Thickness of the Tapeats Sandstone in the Grand Canyon area generally ranges from 0 m to 122 m (Middleton et al., 2003; Billingsley, 2000). It grades from thick beds of cross-bedded and horizontally stratified sandstone, locally pebbly in the lower few meters, upward into thinly layered sandstones interbedded with shales. These strata become younger eastward from late Early to early Middle Cambrian age (Middleton et al., 2003). Facies patterns suggest deposition in intertidal to subtidal shallow-marine environments, with local beach and fluvial deposits (McKee and Resser, 1945; Wannless, 1973; Martin et al., 1986; Elliott and Martin, 1987; Rose et al., 1998; Middleton et al., 2003). Thickness of the Tapeats Sandstone in the Grand Canyon area generally ranges from 0 m to 122 m (Middleton et al., 2003; Billingsley, 2000). These strata are interpreted to have accumulated in an open-shelf environment during eastward marine transgression, with overall sediment derivation from the cratonal platform (McKee and Resser, 1945; Wannless, 1973; Martin et al., 1986; Elliott and Martin, 1987; Rose et al., 1998; Middleton et al., 2003). Thickness of the Bright Angel ranges from 80 m to 150 m.

Muav Limestone (Middle Cambrian)

The Muav consists of cliff-forming limestone, dolomite, and calcareous mudstone that range in thickness from 42 m to 252 m (Middleton et al., 2003; Billingsley, 2000). The Middle Cambrian Muav is interpreted to have

Bright Angel Formation (Lower and Middle Cambrian)

The Bright Angel Formation consists of greenish to brown siltstone and shale interlayered with brownish sandstone (mainly in lower part of unit) and minor gray dolomite (mainly in upper part of unit) (Middleton et al., 2003; Billingsley, 2000). These strata are interpreted to have accumulated in an open-shelf environment during eastward marine transgression, with overall sediment derivation from the cratonal platform (McKee and Resser, 1945; Wannless, 1973; Martin et al., 1986; Elliott and Martin, 1987; Rose et al., 1998; Middleton et al., 2003). Thickness of the Bright Angel ranges from 80 m to 150 m.

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accumulated in subtidal to peritidal environments with deeper-water basins surrounding local offshore shoals (Middleton et al., 2003).

**Temple Butte Formation (Middle and Upper Devonian)**

The Temple Butte Formation consists of limestone, dolomite, mudstone, sandstone, and conglomerate that crop out in a discontinuous layer between the Muav and Redwall limestones (Beus, 2003a). These strata are commonly preserved in channels developed in the underlying Muav Limestone. The Temple Butte Formation is generally interpreted to have formed in narrow tidal channels carved into a shallow, west-facing continental shelf (Beus, 2003a). Thickness ranges from 0 to as much as 140 m.

**Redwall Limestone (Mississippian)**

The Redwall Limestone is a conspicuous, reddish-weathering cliff-forming unit about midlevel in the Grand Canyon stratigraphy (Fig. 1; Beus, 2003b). It ranges from 120 m to 245 m in thickness and rests unconformably on Cambrian (Muav) or, locally, Devonian (Temple Butte) strata. The Redwall formed in a shallow and broad epeiric sea on the western shelf of North America (Beus, 2003b).

**Surprise Canyon Formation (Upper Mississippian)**

The Surprise Canyon Formation consists of reddish-brown siltstone, sandstone, and minor conglomerate that are interlayered with gray limestone (Billingsley and McKee, 1982; Beus, 2003b; Billingsley, 2000). These strata occur as lens-shaped troughs carved into the upper Redwall Limestone, with a maximum thickness of 122 m (Billingsley and Beus, 1985). Most of the paleochannels are oriented east-west with paleoflow toward the west (Glover, 1987; Beus, 2003b).

**Supai Group (Lower Pennsylvanian through Lower Permian)**

The Supai Group consists of a highly variable assemblage of clastic and carbonate rocks that accumulated on a broad coastal plain during early Pennsylvanian (or possibly latest Mississippian) through Early Permian time (McKee, 1982; Blakey, 2003). The primary characteristics of the unit is interlayering of cliff-forming sandstones and limestones with slope-forming siltstones and mudstones. Bedding thickness is highly variable. These strata are interpreted to have formed in a range of fluvial, shallow-marine, and coastal eolian dune settings that were highly migratory due to changing sea level and uplift of the Ancestral Rocky Mountains (Blakey, 2003, 2008, 2009a, 2009b). The Supai Group is divided into the following four formations.

**Watahomigi Formation (Lower Pennsylvanian)**

The Watahomigi Formation consists mainly of reddish siltstone and mudstone interlayered with gray limestone and dolomite that rest unconformably on the underlying Redwall Limestone (Blakey, 2003). The unit thickens westward across the Grand Canyon area from 30 m to 90 m.

**Manacacha Formation (Middle Pennsylvanian)**

The Manacacha Formation consists mainly of quartz-rich sandstone that reaches a maximum thickness of 90 m in the central Grand Canyon (Blakey, 2003). Cross-beds throughout the eolian sandstones consistently record transport from the north.

**Wescogame Formation (Upper Pennsylvanian)**

In contrast to the underlying sandstones of the Manacacha Formation, the Wescogame Formation consists largely of slope-forming siltstone and mudstone with minor sandstone and limestone layers (Blakey, 2003; Billingsley, 2000). These strata are 30–60 m in thickness.

**Esplanade Sandstone (Lower Permian)**

The Esplanade is a dominant cliff-former in the Grand Canyon, with a thickness that increases westward from 75 to 240 m (Blakey, 2003). Most strata are quartz-rich sandstones with abundant cross-bedding that reflects derivation of sediment from the north.

**Hermit Formation (Lower Permian)**

The Hermit Formation is a slope-forming unit consisting mainly of reddish siltstone, mudstone, and fine-grained sandstone (Billingsley, 2000; Blakey, 2003). Thickness is highly variable, ranging from 30 m to over 270 m. Depositional environments were mainly fluvial in a broad floodplain, with minor eolian input (Blakey, 2003).

**Coconino Sandstone (Lower Permian)**

The tall white cliffs of cross-bedded Coconino Sandstone are one of the hallmarks of the Grand Canyon (Fig. 1). These strata range from 20 m to 180 m in thickness and consist largely of quartz-rich fine- to medium-grained sandstone (Middleton et al., 2003). As noted by these authors, the orientation of cross-beds records paleotransport from the north.

**Toroweap Formation (Lower Permian)**

The Toroweap Formation crops out as a thin but laterally continuous slope-forming unit between cliffs of the underlying Coconino Sandstone and overlying Kaibab Limestone (Fig. 2). Dominant lithologies include grayish siltstone and sandstone commonly interlayered with gypsumiferous beds with prominent layer of gray limestone in the middle of the unit (Turner, 2003; Billingsley, 2000). Depositional environments were transitional shallow marine and evaporitic, with influx of eolian sands from the northeast (Turner, 2003). Average thickness is ~100 m (Billingsley, 2000).

**Kaibab Limestone (Lower Permian)**

The Kaibab Limestone forms the prominent upper ledge of the Grand Canyon (Fig. 1). It consists of a lower unit (Fossil Mountain Member) of cliff-forming thin- to medium-beded limestone overlain by a highly variable, slope-forming sequence (Harrsiburgh Member) of limestone, siltstone, sandstone, and gypsum (Hopkins and Thomson, 2003). Hopkins and Thomson interpreted the Kaibab as having accumulated in shallow-marine environments on a broad westward-facing shelf. Average thickness is ~150 m (Billingsley, 2000).

**GEOCHRONOLOGIC SAMPLES AND RESULTS**

**Tapeats Sandstone (Lower and Middle Cambrian)**

Two samples were analyzed from the Tapeats Sandstone, one from near the base of the formation (Tapeats 1) and one from near the top (Tapeats 2) (Fig. 2). The lower sandstone is coarse grained and feldspathic, with pebbles of schist and granite, whereas the upper sample consists of well-sorted, medium-grained, quartz-rich sandstone. Zircons in both samples are mainly colorless to light pinkish, with low degrees of rounding and sphericity. Grain size is variable, with zircons up to
~250 µm in length in the lower unit but only up to ~150 µm in length in the upper unit.

The two samples yield very similar age distributions (Figs. 4 and 5), with two main age groups of ca. 1.45 Ga and ca. 170–1.75 Ga (age peaks of 1455 and 1711 Ma for sample 1 and 1450 and 1736 Ma for sample 2). A K-S comparison of the two samples (Table DR1 [see footnote 1]) yields a P value of 0.99, which demonstrates that the two samples have very similar age distributions. This is somewhat surprising given that the lower sample is part of the basal conglomerate, resting directly on Precambrian basement, whereas the upper unit likely formed in a more integrated drainage system.

**Bright Angel Formation (Lower and Middle Cambrian)**

Three samples of fine-grained sandstone were collected from ~1 m above the Tapeats Sandstone (lower Bright Angel) and from ~5 m (Bright Angel) and ~3 m (upper Bright Angel) below the top of the unit (Fig. 2). Zircons in the samples are very small (<100 µm in length), euhedral to only slightly rounded, and generally colorless to light pinkish. Analyses were conducted with a laser beam diameter that ranged from 10 to 18 µm. The three samples yield very similar ages, with main age groups of ca. 1.03, 1.45, and 1.71 Ga (Fig. 5). With all three samples combined, primary age peaks are at 1029 Ma (n = 12), 1457 Ma (n = 63), and 1712 Ma (n = 113) (Fig. 6; Table DR1 [see footnote 1]). K-S analysis P values for comparison of the three samples range from 0.23 and 0.74 (Table DR1 [see footnote 1]).

**Temple Butte Formation (Middle and Upper Devonian)**

Three samples were collected from the Temple Butte Formation (Fig. 2). Two of the samples are medium-grained sandstones (Temple Butte 1 and 3), whereas sample Temple Butte 2 is a pebbly, coarse-grained sandstone. Zircons from the three samples are generally larger in the conglomeratic sample (up to ~150 µm in length). 

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Figure 6. Diagram showing the distribution of detrital zircon ages from each unit. Each curve contains all analyses from a single unit, normalized such that all curves contain the same area. The number of constituent analyses for each curve is shown on the left. The youngest age of deposition of each unit (vertical lines) is based on the depositional ages reported in Beus and Morales (2003) and the Ogg et al. (2008) time scale. Also shown are age distributions for strata of the underlying Unkar and Chuar Groups in the Grand Canyon (from Timmons et al., 2005; Bloch et al., 2006; K. Karlstrom, 2010 written commun.). Vertical shaded bars show the main ages of zircons that would have been shed from various potential source regions. These age ranges have been compiled primarily from Hoffman (1989) and Dickinson and Gehrels (2008a). The cumulative probability plots provide an additional graphical distinction among Cambrian–Devonian, Mississippian–Pennsylvanian–lowest Permian, and lower Permian units.
length in samples 1 and 3, up to ~250 µm in length in sample 2), colorless to light pink, and poorly rounded.

All three samples yield similar age distributions, with primary age peaks of ca. 1.44 Ga and ca. 1.65–1.75 Ga and subordinate ages of ca. 400–600 Ma (Fig. 5). K-S analysis P values indicate that samples 2 and 3 are indistinguishable (P value = 1.0), whereas sample 1 is somewhat less similar due to a lower proportion of ca. 1.44 Ga grains (P values = 0.20/0.09 with TB2/TB3). With all three samples combined (302 analyses; Fig. 6), peak ages are 518 Ma (n = 8), 1122 Ma (n = 6), 1442 Ma (n = 75), 1740 Ma (n = 150), 1900 Ma (n = 29), and 2069 Ma (n = 5).

Detrital zircon ages in the Temple Butte Formation are quite similar to ages in the underlying Bright Angel and Tapeats sandstones, with K-S analysis P values of 0.39 (with Tapeats) and 0.41 (with Bright Angel).

**Surprise Canyon Formation (Mississippian)**

Two samples of sandstone and one sample of conglomeratic sandstone were analyzed from this unit. Zircons in the samples are mainly colorless to light pink, less than ~100 µm in length, and not well rounded.

Ages from two of the samples (1 and 2) are quite similar, with a K-S analysis P value of 0.65. The third sample contains similar age groups, but in very different proportions, and accordingly has a P value of 0.0 in comparison with the other two samples. A composite age distribution of the three samples reveals dominant age peaks of 416 Ma (n = 14), 473 Ma (n = 9), 1030 Ma (n = 59), 1066 Ma (n = 62), 1191 Ma (n = 40), 1366 Ma (n = 20), 1442 Ma (n = 23), 1679 Ma (n = 15), and 1755 Ma (n = 12) (Fig. 6; Table DR1 [see footnote 1]).

The ages from these samples are quite different from ages in underlying units, with significant age groups in the early Paleozoic (400–500 Ma) and mid-Proterozoic (1.0–1.2 Ga), and reduced proportions of 1.40–1.46 Ga and 1.6–1.8 Ga ages (Figs. 5 and 6). The proportions of 1.62–1.70 Ga and 1.70–1.80 Ga ages are also different, with 57% younger grains (age peak at 1677 Ma), and 43% older grains (age peak at 1732 Ma) (Fig. 6; Table DR1 [see footnote 1]).

**Supai Group**

**Watahomigi Formation (Lower Pennsylvanian)**

Our sample of Watahomigi Formation was collected from a reddish, fine-grained sandstone in the lowest sandstone layer, ~5 m above the Redwall Limestone. Zircons from the sample are small to medium in size (up to ~150 µm in length), variable in color (from colorless to medium pinkish in color), and slightly rounded. The main ages of zircons are 1.6–1.8 Ga (n = 31), with significant age groups of 1.8–3.0 Ga (n = 25), 1.0–1.2 Ga (n = 18), and 390–620 Ma (n = 10) (Figs. 4 and 5).

**Manacacha Formation (Middle Pennsylvanian)**

A sample of medium-grained sandstone was collected from one of the thicker and more resistant sandstone layers in the lower part of the Manacacha Formation. Zircons from the sample are generally colorless to medium pink, large (up to ~250 µm in length), moderately to well rounded, and moderately spherical. The resulting ages are mostly Precambrian, with only two ages between 400 and 450 Ma, 24 ages between 1.0 and 1.2 Ga, 32 ages between 1.6 and 1.8 Ga, and 22 ages between 1.8 and 3.0 Ga (Figs. 4 and 5).

**Wescogame Formation (Upper Pennsylvanian)**

One sample of Wescogame Formation was collected from a medium-grained sandstone. Zircons in the unit are mainly colorless to light pink, up to ~200 µm in length, and euhedral to slightly rounded. Main age groups are 450–470 Ma (n = 2), 650–670 Ma (n = 2), 1.0–1.2 Ga (n = 19), and 1.6–1.8 Ga (n = 24). There are also 27 ages in the 1.8–3.0 Ga range (Figs. 4 and 5).

**Esplanade Sandstone (Lower Permian)**

Three samples were collected from medium-grained sandstones in the middle, upper, and uppermost Esplanade Formation (Fig. 2). Zircons in all three samples are colorless to medium pink, of moderate size (up to ~200 µm in length), and are moderately rounded/spherical. These samples yield age spectra with main age groups of 330–480 Ma (n = 15), 660–700 Ma (n = 3), 1.0–1.2 Ga (n = 70), 1.6–1.8 Ga (n = 92), and 1.8–3.0 Ga (n = 57) (Figs. 4 and 5).

**Summary of Supai Ages**

Detrital zircon ages from six samples of Supai strata are quite similar (Fig. 5), with K-S analysis P values that are all ≥0.05 (Table DR1 [see footnote 1]). The six sets of ages are accordingly combined on Figure 6 into two age distributions of Pennsylvanian age and earliest Permian (Wolfcampian) age. These two age distributions are similar (K-S P value of 0.41; Table DR2 [see footnote 1]), with main groupings of early Paleozoic, mid-Proterozoic, and Early Proterozoic ages (Fig. 6). There are also several grains in each sample set between 600 and 720 Ma.

As shown on Figure 6, the detrital zircon ages in the Supai Group are very similar to those in the Surprise Canyon Formation. The proportions of ages are somewhat different, however, in that there is a higher proportion of Early Proterozoic ages in the younger strata. This results in significantly different cumulative probability curves for the Surprise Canyon versus Supai sample sets (Fig. 6, upper inset), and a P value of 0.0 (Table DR2 [see footnote 1]).

**Hermit Formation (Lower Permian)**

We collected three samples of fine-grained sandstone, one sample from near the base of the formation and two from near the top (Fig. 2). Zircons in all three samples are generally <80 µm in size, well rounded, and colorless to light pinkish in color. Because of the small grain size, analyses were conducted with a laser beam diameter that ranged from 10 to 18 µm. Dominant age groups are generally similar in the three samples, with dominant groups of mid- and Early Proterozoic and an extended range of Late Proterozoic–Early Paleozoic ages (Fig. 5). The proportion of these youngest ages also tends to increase up section, with younger than 720 Ma grains making up 10% of the population in the lower sample, 21% in the middle sample, and 37% in the upper sample. This change in proportions is consistent with K-S P values of 0.51 and 0.09 for comparisons of the middle with upper and lower samples, but only 0.003 for comparison of the upper and lower samples. The combined curve shows a profound change in the distribution of young ages compared to the underlying Esplanade Sandstone (Fig. 6), specifically because of the occurrence of a significant number of grains having ages between 500 and 720 Ma (age peaks at 527, 580, and 630 Ma).

**Coconino Sandstone (Lower Permian)**

Two samples were collected from medium-grained sandstones in the lower and upper portions of the Coconino Sandstone (Fig. 2). Zircons in both samples are generally colorless to dark pink in color, large (grains up to ~250 µm in length), and generally well rounded and moderately spherical. Zircons from sample 1 were initially analyzed and reported by Dickinson and Gehrels (2003), but they were reanalyzed herein to take advantage of improvements in analytical precision. The age distributions in the two samples contain similar proportions of similar ages (Fig. 5), resulting in a high P value (0.99; Table DR1 [see footnote 1]). With the two age distributions combined, the dominant ages are 320–700 Ma (n = 26), 1.0–1.2 Ga (n = 63), 1.6–1.8 Ga (n = 39), and 1.8–3.7 Ga (n = 32) (Fig. 6).
Toroweap Formation (Lower Permian)

Two samples were collected from fine- to medium-grained sandstone layers in the upper portion of the Toroweap Formation (Fig. 2). Zircons in the samples are highly variable in size (up to 200 µm in length), color (ranging from colorless to dark pink), and degree of rounding (ranging from euhedral crystals to well-rounded and highly spherical grains). The samples yield similar age groups, but in somewhat different proportions, as indicated by a K-S \(^2\) value of 0.04 (Table DR2 [see footnote 1]). Based on a combined age distribution, primary ages are 300–700 Ma \((n = 45)\), 1.0–1.2 Ga \((n = 34)\), 1.6–1.8 Ga \((n = 48)\), and 1.8–3.0 Ga \((n = 29)\) (Fig. 6).

Kaibab Formation (Lower Permian)

We analyzed samples from the lowest and highest sandstones observed in the Kaibab Formation (Fig. 2). Both are thin sandstone layers interbedded with gray limestone beds in the upper part of the formation (Fig. 2). Zircons in the samples are small (<100 µm), variably rounded, and mainly colorless to light pink in color. The resulting ages are similar in the two units, but occur in different proportions, resulting in a K-S \(^2\) value of 0.03 (Table DR1 [see footnote 1]). With combined ages, the main age groups are 270–360 Ma \((n = 12)\), 380–500 Ma \((n = 15)\), 530–740 Ma \((n = 17)\), 1.0–1.3 Ga \((n = 55)\), 1.40–1.52 Ga \((n = 23)\), and 1.6–1.8 Ga \((n = 38)\).

PROVENANCE INTERPRETATIONS

Provenance of the detrital zircons is discussed primarily in relation to the age provinces of southern North America, as summarized in terms of geography on Figure 7 and in terms of age with vertical color bands on Figure 6. The provenance of each set of units is discussed separately next.

Tapeats Sandstone (Lower and Middle Cambrian)

The grains in these samples are very similar to ages of crystalline basement rocks throughout the southwestern United States, with dominantly Early Proterozoic grains (peak age = 1728 Ma; \(n = 102\)) and a lower proportion of 1.4–1.5 Ga grains (peak age = 1452 Ma; \(n = 65\)). In fact, 69% of the Early Proterozoic grains are interpreted to have been shed from the 1.70–1.80 Ga (Yavapai) province, with 31% derived from the 1.62–1.70 Ga (Mazatzal) province (Fig. 7; Karlstrom et al., 2003). There are also a few Grenville-age (1.0–1.2 Ga) grains, with an age peak at 1064 Ma \((n = 8)\). These grains were presumably shed from the Grenville orogen to the south and/or east (Fig. 7).

Comparison of these ages with ages in Proterozoic strata of the locally underlying Grand Canyon Supergroup (Unkar and Chuar Groups; Fig. 6) suggests that little detritus was recycled from underlying units.

Bright Angel Formation (Lower and Middle Cambrian)

Ages from the Bright Angel Formation are very similar to those in the underlying Tapeats Sandstone (Fig. 6), with a K-S \(^2\) value of 0.37 in comparison of the two sets of samples (Table DR2 [see footnote 1]). Provenance interpretations are accordingly similar; most detritus was shed from southwestern United States crystalline basement and from the Grenville orogen to the south or east. The slightly greater proportion of 1.62–1.70 Ga grains (35%) over 1.70–1.80 Ga grains (65%) may reflect cratonward migration of source regions during Cambrian transgression.

Temple Butte

The Temple Butte Formation yields Paleo-proterozoic and Mesoproterozoic ages that are similar to those in the Tapeats and Bright Angel

Figure 7. Location of the main age provinces in central and southern North America that may have provided sediment for Paleozoic strata of the Grand Canyon. Figure is adapted from Anderson and Morrison (1992), Bickford et al. (1986), Hoffman (1989), Burchfiel et al. (1992), Bickford and Anderson (1993), Van Schmus et al. (1993), Dickinson and Lawton (2001), and Dickinson and Gehrels (2009a).
Formations, even with similar proportions of 1.62–1.70 Ga (30%) versus 1.70–1.80 Ga (70%) grains (Fig. 6). Most of the detritus is accordingly interpreted to also have been shed from Precambrian basement rocks of the southwestern United States.

Additional sources are recorded by grains of 1.34–1.40 Ga (age peak at 1376 Ma; \( n = 8 \)) in all three samples, and grains of 403 and 426 Ma in sample 1. The grains with an age peak at ca. 1376 Ma were most likely derived from the 1.34–1.40 Ga midcontinent granite-rhyolite province of the south-central United States (Anderson and Morrison, 1992; Van Schmus et al., 1993; Bickford and Anderson, 1993) (Fig. 7). The ca. 518 Ma grains may have been shed from the Antler orogen in southeastern Nevada (Gehrels and Dickinson, 2000); Devonian foreland basin strata of the Appalachian (Gray and Zeitler, 1997; McLennan et al., 2001; Eriksson et al., 2004; Thomas et al., 2004; Becker et al., 2005, 2006; Park et al., 2010).

An alternative interpretation is that the ca. 518 Ma grains, as well as the single 403 Ma and 426 Ma grains, were shed from circum–North American orogenic belts. Specifically, Park et al. (2010) reported similar ages from Devonian foreland basin strata of the Appalachians. This interpretation is also problematic in that early Paleozoic detrital zircons shed from circum–North American orogenic belts are always accompanied by a significant (generally greater) proportion of Grenville-age grains. The possibility of derivation from circum–North American orogenic systems is described in more detail later herein because younger units contain abundant Paleozoic and mid-Proterozoic zircons.

**Surprise Canyon Formation (Mississippian)**

The occurrence of significant proportions of younger than 1.3 Ga zircons in the Surprise Canyon Formation and younger strata (Fig. 6) signals a major change in the provenance of sandstones in the Grand Canyon. Possible sources for these grains include the Cordilleran orogen to the west, Franklinian (Innuitian-Ellesmere) orogen to the north, Ouachita orogen to the south, and Appalachian orogen to the east. Fortunately, all of these options can be evaluated in at least a preliminary fashion with existing detrital zircon data. Figure 8 presents these comparisons with age-distribution curves that are upward facing for Mississippian–Permian strata of the Grand Canyon and downward-facing for Upper Paleozoic strata of other regions. All curves are normalized for the number of constituent analyses.
Derivation from the Antler orogen in the western United States is a possibility given that clastic detritus was shed eastward from the Antler Highlands (Fig. 7), onto the North American craton, during Mississippian and Pennsylvanian time (Burchfield et al., 1992). Detrital zircons in Mississippian strata that accumulated within and east of the highlands yield only Precambrian ages (Gehrels and Dickinson, 2000) and are a good match for rocks exposed in the Antler Highlands (Gehrels et al., 2000). However, because there is little overlap of these ages with the detrital zircon ages in Upper Paleozoic strata of the Grand Canyon, the Antler orogen is not a likely source for the Grand Canyon strata (Fig. 8).

Derivation from the north is also possible given that tectonism, referred to as the Franklinian, Innuitian, and Ellesmere orogenic events, occurred along the Arctic margin during mid- and late Paleozoic time (Trettin, 1989). A regionally extensive clastic wedge was shed southward, and probably covered much of the Canadian Shield during mid- and late Paleozoic time (Patchett et al., 2004). Detrital zircon information is available from Devonian strata belonging to this clastic wedge (McNicoll et al., 1995), from Devonian–Mississippian strata of eastern Alaska (Nation River Formation; Gehrels et al., 1999) and northwestern Canada (Beranek et al., 2010) that are known or interpreted to have been part of this clastic wedge, and from Cretaceous strata that rest on, and were at least in part recycled from, Franklinian clastic strata (Røhr et al., 2010). As reported by previous workers, most of this detritus is interpreted to have been shed from the northern continuation of the Appalachian-Caledonian orogenic system. The similarity of the Franklinian curve with the Grand Canyon age distribution (Fig. 8B) suggests that detrital zircons in the Grand Canyon strata could have been shed from the Franklinian orogen, although ultimately, they would have originated in the northern continuation of the Appalachian-Caledonian orogen.

Evaluation of the Ouachita orogen is enabled by comparison with Pennsylvanian strata of the Jackfork Formation in the Ouachita Mountains (Arkansas) and the Hayden Formation of the Marathon basin (Texas) (Gleason et al., 2007) (Fig. 8C). These units yield Mesoproterozoic to Paleoproterozoic ages and 350–500 Ma ages that match with the Grand Canyon strata, but they also reveal a significant population of ca. 520 Ma grains. The scarcity of these ca. 520 Ma ages in Mississippian Grand Canyon strata suggests that the Ouachita orogen is not a likely source terrane, although this conclusion needs to be reevaluated when additional detrital zircon data from the Ouachita system become available.

Derivation from the Appalachian orogen can be evaluated by comparison with abundant detrital zircon data from Appalachian foreland basin strata reported by Gray and Zeitler (1997), McLennan et al. (2001), Eriksson et al. (2004), Thomas et al. (2004), Becker et al. (2005, 2006), and Park et al. (2010). As shown on Figure 8D, the composite age distribution from these Devonian through Permain strata provides an excellent match for the younger than 1.4 Ga Grand Canyon ages. Figure 9 provides a more detailed comparison of detrital zircon age distributions from strata of the Grand Canyon (upward-facing curves) and Appalachian orogen (downward-facing curves). All curves are normalized, such that they contain the same area. Dashed lines represent the residual age distribution for ages that could not have been shed from the Appalachian orogen (calculated by subtracting the Appalachian age distribution from the Grand Canyon age distribution). The area beneath the dashed line compared to the area beneath the full Grand Canyon curve is indicated as a percent for each age range. Vertical shaded bars show the main ages of zircons that would have been shed from various potential source regions. These age ranges have been compiled primarily from Hoffman (1989) and Dickinson and Gehrels (2009a). Comparisons are provided for (A) Cambrian through Silurian strata (residual age peaks of 1451 and 1721 Ma), (B) Devonian strata (residual age peaks at 1433 and 1737 Ma), (C) Mississippian strata (residual age peaks at 1435, 1678, and 1740 Ma), (D) Pennsylvanian and lowest Permian strata (residual age peaks at 1539, 1657, and 1781 Ma), (E) Permain strata (residual age peaks at 426, 526, 520, 617, 1426, 1510, 1653, and 1765 Ma).
analysis by sequentially comparing ages of generally coeval strata from the Grand Canyon and the Appalachian orogen. As with Figure 8, Grand Canyon age distributions are shown with upward-facing curves, while Appalachian ages are shown with downward-facing probability curves. All curves are normalized for the number of constituent analyses.

Mismatches between coupled curves are shown with dashed lines for each time period (Fig. 9), and these can be interpreted as the age distribution of zircons in each set of Grand Canyon samples that could not have been derived from the Appalachian orogen. These dashed lines may also serve as an indication of the proportion of zircon derived from non-Appalachian sources, given that the age-probability curves are normalized. Following this interpretation, values representing the area beneath each dashed line divided by the area beneath the full Grand Canyon age-distribution curve are shown for each set of units (Fig. 9). In evaluating this analysis of proportions of detrital zircon grains, it is important to realize that variations in zircon fertility complicate a simple conversion from abundance of detrital zircons to relative volumes of rock material (Dickinson, 2008).

Based on this analysis, we suggest that clastic strata of the Surprise Canyon Formation consist primarily (~75%) of zircons derived from the Appalachian orogen, with subordinate (~25%) zircons derived from Precambrian basement rocks. The abundance of Mazatzal over Yavapai grains (Fig. 6) and the scarcity of detritus older than 1.8 Ga suggest that the river systems responsible for carrying this material across the craton did not flow across the Canadian Shield to the north (Fig. 7), or that the craton in northern and perhaps central North America remained covered during this time (Patchett et al., 2004). Exposure of older than 1.4 Ga source rocks may have occurred during early phases of uplift of the Ancestral Rocky Mountains (Burchfiel et al., 1992).

**Supai Group**

The Supai Group contains similar age groups as the Surprise Canyon Formation, with older grains that are an excellent match for basement rocks of the southwestern United States and younger grains that are an excellent match with the Appalachian orogen (Fig. 6). Figure 9D provides a specific comparison of ages in Pennsylvanian–lowest Permian strata of the Supai Group with ages in Pennsylvania foreland basin strata of the Appalachians. As with the Mississippian comparison, the similarities are impressive, suggesting that the Appalachian orogen is a viable source region for much of the detritus in the Supai Group.

The difference between the Supai and Appalachian curves (Fig. 9D) leaves a residual age distribution (dashed line) with dominant age peaks of 1539 Ma, 1657 Ma, and 1781 Ma, and an area of 39% of the full Grand Canyon age-distribution curve. The older grains (e.g., 1657 Ma and 1781 Ma) most likely record derivation from crystalline rocks exposed in the Ancestral Rocky Mountains, which were certainly high and generating sediment during Pennsylvanian–earliest Permian time (e.g., Kluth and Coney, 1981; Blakey et al., 1988; Johansen, 1988; Marzolf, 1988; Peterson, 1988; Burchfiel et al., 1992; Blakey, 2009a). It is interesting to note that the proportion of Mazatzal-Yavapai zircons in Supai strata, 57%–43%, is similar to the underlying Surprise Canyon Formation, suggesting that the earliest Ancestral Rockies may also have been a sediment source during late Mississippian time.

An additional mismatch is represented by the age peak at 1539 Ma (Fig. 9D), which overlaps with age peaks in many of the individuals (Fig. 5) and composite (Fig. 6) age distributions for Upper Paleozoic samples. These ages are of uncertain origin because they do not appear in the Appalachian reference curve and do not match ages of igneous rocks on the North American craton. Possible sources for these grains are discussed later herein.

**Hermit, Coconino, Toroweap, and Kaibab Formations**

As shown on Figure 6, the age distributions from the Hermit, Coconino, Toroweap, and Kaibab Formations are all quite similar to each other and, taken together, different from underlying units. These similarities are indicated by K-S P values of 0.10–0.91 for all interunit comparisons (Table DR2 [see footnote 1]), and the differences with underlying strata are indicated by low P values (0.002 or lower) in comparisons with all underlying units except for the Surprise Canyon Formation (P values up to 0.18).

Although a major change in age distribution is noted between the composite curves for the Hermit Formation and Supai Group, the increasing proportion of young zircons and the appearance of 270–380 Ma and 480–720 Ma populations within the Hermit Formation suggest that the change occurs within the Hermit unit rather than along its base. We accordingly infer that a dramatic change in provenance occurred during early Leonardian time (age of the Hermit; Blakey, 2003).

In terms of provenance, the detrital zircon ages in the Hermit through Kaibab Formations are a reasonably good match for Permian foreland basin strata of the Appalachian orogen, which have been reported by Becker et al. (2006) (Fig. 9E). Comparison of the two age distributions leaves a residual age probability of 44% of the full Grand Canyon curve, with significant numbers of grains between 420 and 720 Ma (age peaks at 426, 526, 578, and 617 Ma), and at 1426, 1510, 1653, and 1765 Ma. As with underlying units, the ca. 1.4 and 1.6–1.8 Ga grains could well have been shed from basement rocks of the central or southwestern United States. The proportions of Mazatzal versus Yavapai ages are 59% and 41%, which are very similar to proportions for the Supai Group and Surprise Canyon Formation. The chronology of a major change in provenance during accumulation of the Hermit Formation (early Leonardian according to Blakey [2003], or ca. 275 Ma according to Ogg et al. [2008]) is an excellent match for the timing of final emplacement of outboard terranes and final collision with western Africa along the southern Appalachian margin (Dallmeyer, 1989; Hatcher et al., 1989).

The main mismatch is in ages of 420–720 Ma, which are present but in low abundance in the Appalachian Permian reference curve (Fig. 9). We consider three possible reasons for this mismatch in ages. First, there is the possibility that the Permian strata analyzed by Becker et al. (2006), the Dunkard Group, predate the Hermit and younger strata and accordingly accumulated prior to the early Leonardian change in provenance. Unfortunately, the biostratigraphic control on the Dunkard Group is not adequate to evaluate this possibility. A second possibility is that the 420–720 Ma grains were shed mainly from the southern Appalachians, and are accordingly under-represented in Becker et al.’s (2006) samples from Ohio and Pennsylvania. This is unlikely, however, because igneous rocks of the appropriate ages are also present in the central Appalachians. These igneous rocks include rift assemblages emplaced along the North American margin (Tollo et al., 2004) and igneous rocks in the Avalon, Carolina, and Suvané terranes, which were involved in the final collision of Africa and North America during Alleghanian tectonism (Heatherington et al., 1996; Wortman et al., 2000; Hibbard et al., 2002; Dickinson and Gehrels, 2009a; Park et al., 2010). A third possibility is that the 420–720 Ma grains were shed from the Ouachita orogen, rather than the Appalachians. This possibility cannot be tested directly with the available detrital zircon data from the Ouachita orogen (only strata of Pennsylvanian age have been studied), but it is consistent with the occurrence of Neoproterozoic igneous rocks of appropriate ages in the Yucatan-Campeche blocks (as
SEDIMENTARY LAG TIMES

The ages of the youngest detrital zircon in each sample can be compared with the depositional age of the host strata to constrain the duration between crystallization of zircons in source terranes and accumulation of the host sediment. This comparison is shown on Figure 6, where the age-distribution curves record the youngest ages in each set of samples, and thick vertical lines represent interpreted ages of deposition. The latter are based on the depositional ages reported in Beus and Morales (2003), which have been converted to absolute age using the Ogg et al. (2008) time scale. The youngest grains are reported as the youngest single grain, which is to be used with caution because of possible Pb loss, and also the youngest peak in age probability with at least three constituent ages, which is a more robust indicator of maximum depositional age (from Table DR1; methodology from Dickinson and Gehrels, 2009b). The patterns are as follows:

1. The Tapeats Formation yields ages of 1006 Ma (youngest grain) and 1064 Ma (youngest multigrain peak), which are considerably older than the ca. 499 Ma (late Middle Cambrian or older; Middleton et al., 2003) age of deposition.

2. The Bright Angel Formation reveals youngest ages of 757 and 1029 Ma, which are also significantly older than the depositional age of ca. 499 Ma (late Middle Cambrian or older; Middleton et al., 2003).

3. The Temple Butte Formation yields ages of 403 and 517 Ma, compared with a depositional age of ca. 375 Ma (early Late and possibly late Middle Devonian; Beus, 2003a).

4. The Surprise Canyon Formation has ages of 309 Ma and 416 Ma, compared with the depositional age of ca. 318 Ma (late Mississippian [Chesterian]; Beus, 2003b).

5. The Supai Group has ages of 333 Ma and 407 Ma, compared with the depositional age range of ca. 313 Ma to ca. 275 Ma (early Pennsylvanian [Morrowan] to Early Permian [Leonardian]; Blaeky, 2003).

6. The Hermit through Kaibab Formations collectively yield youngest ages of 273–329 Ma and multigrain peak ages of 295–433 Ma (Table DR1 [see footnote 1]), compared with a depositional age of ca. 275 to ca. 268 Ma (no younger than latest Early Permian [latest Leonardian]; Hopkins and Thomson, 2003).

The average time lags for the Devonian and younger samples, based on the youngest single-grain ages, range from ~9 m.y. to 58 m.y., with an average of 25 m.y. The youngest multigrain peak ages, which are more robust indicators of maximum depositional age (Dickinson and Gehrels, 2009b), average 95 m.y. Both of these apparent lag times suggest that Devonian through Permian sandstones of the Grand Canyon consist at least in part of detritus that records relatively rapid erosion, transport, and final accumulation. Most of the zircons in these samples significantly predate deposition, however, which raises the possibility of recycling prior to final accumulation.

PALEOGEOGRAPHY RECORDED BY PALEOZOIC STRATA OF THE GRAND CANYON

Paleogeographic implications of the five main phases of sediment provenance recorded by sandstones of the Grand Canyon are outlined next and shown on Figure 10.

Cambrian

The first phase of sediment provenance is recorded by the Tapeats and Bright Angel Formations, which yield ages that are similar to each other but different from ages in most overlying strata. The age peaks of these samples are most consistent with derivation from the underlying Yavapai and Mazatzal Provinces (Karlstrom and Bowring, 1993; Van Schmus et al., 1993) and from 1.4 to 1.5 Ga granitic rocks of the central and western United States (Fig. 7; Anderson and Morrison, 1992). A slight increase in the proportion of 1.62–1.70 Ga ages (31% in the Tapeats, 35% in the Bright Angel) may reflect eastward migration of source regions during Cambrian transgression.

These relations support the facies and paleocurrent patterns reported by McKee and Resser (1945) and summarized by Middleton et al. (2003), which indicate derivation from crystalline basement exposed along the western flank of the Transcontinental Arch, within and to the east of the Grand Canyon region. We accordingly envision a paleogeographic setting for Middle Cambrian time in which sediment was accumulating in a transgressive setting on a west-facing shelf, with deposition grading eastward from subtidal to intertidal to beach and eventually fluvial environments (Fig. 10A).

Devonian

An important change in sediment provenance is recorded by the influx of detrital zircons with ages of ca. 1375 Ma, ca. 518 Ma, and 403–426 Ma in the Middle–Upper Devonian Temple Butte Formation. A reasonable source for most of this detritus is the midcontinent region, where 1.34–1.40 Ga granites and rhyolites as well as ca. 520 Ma granitic rocks (Amarillo-Wichita uplift) are exposed (Fig. 7). These grains are subordinate, however, to 1.4–1.5 Ga and 1.6–1.8 Ga grains that occur in relative proportions similar to the underlying Cambrian strata. The proportion of Mazatzal-Yavapai ages is also similar, with 30% derived from the younger (1.62–1.70 Ga) province. An additional constraint on provenance is based on the lack of Grenville-age (1.0–1.2 Ga) grains in these samples, which precludes sediment derivation from far to the south or east (Fig. 7). Accordingly, we envision a paleogeography during Middle–Late Devonian time in which rivers transported sediment from highlands along the Transcontinental Arch (in the Texas-Oklahoma-New Mexico–Colorado region) that exposed Paleoproterozoic crystalline basement, 1.34–1.48 Ga granitoids and volcanic rocks, and ca. 520 Ma granitoids (Fig. 10B). The occurrence of two grains of 403–426 Ma suggests that some detritus may have come from the Appalachian orogen or perhaps the Franklinian orogen during this time.

Mississippian

The third phase of sediment derivation is recorded by the Mississippian Surprise Canyon Formation. This unit yields ages older than 1.4 Ga that are generally similar to underlying units, with age peaks of 1442 Ma, 1679 Ma, and 1755 Ma. Important differences with older units are that Mazatzal ages are dominant (57%) over Yavapai ages (43%), and that there are few 1.34–1.40 Ga ages in the Surprise Canyon Formation. This suggests that the older grains in the Surprise Canyon were not recycled from...
Figure 10. Schematic depiction of interpreted paleogeography and sediment provenance during accumulation of Cambrian through Permian sandstones of the Grand Canyon. See text for explanation.
underlying strata but instead may have been shed directly from Precambrian basement of the southwestern or midcontinent United States.

The main difference with older strata is the dominance of younger than 1.4 Ga ages in the Surprise Canyon Formation (Fig. 6). The Grenville orogen was presumably the main contributor of zircons, although the proportion of sand of this age was considerably less because of the high zircon fertility of Grenville-age rocks (Moecher and Samson, 2006; Dickinson, 2008). There are also numerous ages between 380 and 480 Ma (Fig. 6). The most likely source for grains of these ages is the Appalachian orogen (Fig. 10C), which contains widespread Ordovician–Devonian igneous rocks that were generated during Taconic and Acadian tectonism (Hatcher, 1989; Hatcher et al., 1989; Osberg et al., 1989; Park et al., 2010). Uplift recorded by the Surprise Canyon Formation would have occurred during the Alleghanian orogeny, which affected primarily the southern Appalachians (Hatcher, 1989; Hatcher et al., 1989).

We accordingly envision a paleogeography for Mississippian time in which major river systems were carrying sediment westward across the North American continent from the Southern and possibly central Appalachians (Fig. 10C). These rivers also incorporated detritus from rocks of 1.6–1.8 Ga age (mainly the Mazatzal Province) and from ca. 1435 Ma granitic bodies that were exposed in the central to southwestern United States.

Pennsylvanian–Earliest Permian

Strata of the Supai Group record a fourth phase in sediment provenance during Pennsylvanian–earliest Permian time. In comparison with older time periods, Supai strata record derivation from the Mazatzal–Yavapai Provinces in identical relative proportions (57% Mazatzal and 43% Yavapai) but greater overall proportion than the Surprise Canyon Formation, few zircons of 1.4–1.5 Ga, and a lesser proportion of Grenville-age zircons. The younger (<1.4 Ga) zircons yield ages that are very similar to the Surprise Canyon Formation, with probable sources in the Appalachian orogen.

We envision a paleogeography for this time period in which large river systems were carrying sediment westward across the continent from the Appalachian orogen, with northeasterly trade winds (e.g., Parrish and Peterson, 1988) transporting and reworking the sediment into local eolian units (e.g., Espilane Formation) (Fig. 10D). These rivers were also incorporating sediment (~39% of the total zircon grains) from Precambrian basement exposed in the Ancestral Rocky Mountains, as suggested by Peterson (1988), Blakey (1988), Blakey et al. (1988), Marzolf (1988), Johansen (1988), and many others. Given the predominance of Mazatzal-age over Yavapai-age grains, the most likely course of these rivers was across the south-central United States, where Ancestral Rocky Mountain uplifts would have exposed mainly Mazatzal Province basement and perhaps a smaller proportion of 1.4–1.5 Ga granitic material (Fig. 7).

Early Permian

The final phase recorded in Grand Canyon stratigraphy is Early Permian (post-Wolfcampian) in age, as recorded by detrital zircon ages from the Hermit, Coconino, Toroweap, and Kaibab Formations. These strata yield similar detrital zircon age populations that are quite distinct from underlying units. As with the Supai Group and Surprise Canyon Formation, most detritus was shed from Paleoproterozoic (mainly Mazatzal) and Grenville-age provinces (Fig. 10F). In contrast to the Supai Group, 1.4–1.5 Ga zircons are abundant, perhaps reflecting a slightly different source within the Ancestral Rocky Mountains (Fig. 8). The main difference with older strata is the larger proportion and greater age range of Neoproterozoic–Paleozoic zircons. These ages are an excellent match for igneous rocks that were generated during later stages of the Alleghanian orogeny.

We accordingly offer a paleogeographic scenario for Permian time (Fig. 10F) that is similar to Pennsylvanian time (Figs. 10D–10E), with large rivers bringing detritus across the continent and northeasterly trade winds (e.g., Parrish and Peterson, 1988) reworking this material into widespread eolian units (e.g., Coconino Sandstone).

SUMMARY

U-Pb analyses of detrital zircons from Paleozoic strata of the Grand Canyon demonstrate that the provenance of sandstones varied significantly from Cambrian through Permian time. As has been recognized by all previous workers, sandstones of the Grand Canyon were derived in large part from Precambrian basement rocks in the southwestern United States. In lower strata, as described by McKee and Resser (1945), nearly all of the sediment appears to have been shed from basement rocks exposed in the Grand Canyon area and in cratonic rocks to the east. In younger strata, as described by Peterson (1988), Blakey (1988), Blakey et al. (1988), Marzolf (1988), Johansen (1988), and many others, an abundance of sand was derived from Precambrian basement exposed in the Ancestral Rocky Mountains to the east.

Our detrital zircon data also demonstrate that much of the sandstone in the upper part of the Grand Canyon section was shed from distant sources, including mid-Proterozoic rocks of the Grenville orogen and Neoproterozoic–Paleozoic rocks in circum–North American orogenic belts. This result was first presented by Dickinson and Gehrels (2003) based on analysis of one sample of Coconino Sandstone, with the conclusion that much of the detritus was derived from the Appalachian orogen. Our more comprehensive data set presented here indicates that all of the sandstones above the Mississippian Redwall Limestone probably originated in large part from the Appalachian orogen.

The conclusion that sediment was transported from the Appalachian orogen to the Grand Canyon region from Mississippian through Permian time has important implications for the paleogeography of North America during late Paleozoic time. The Appalachian region was the continental divide during much of late Paleozoic time and spawned major rivers that flowed northwest. The northern Appalachian and Caledonian mountains may have fed rivers that drained across central Canada into the west coast of North America (Figs. 10B–10F). Such rivers would have tapped sources that could have supplied the zircon age populations in Grand Canyon sedimentary rocks at 625 Ma, 570 Ma, 424 Ma, and 412 Ma. However, other paleogeographic elements complicated the paths of fluvial systems that drained the central and southern Appalachian systems. The Grand Canyon age peaks at 384 Ma, 352 Ma, 338 Ma, and 310 Ma were likely derived from farther south in the Appalachian orogen. Throughout Paleozoic time, at least until the middle Pennsylvanian, the Transcontinental Arch prevented direct flowage of southern Appalachian–derived rivers into the Western Interior (Figs. 10A–10D; see also Blakey, 2009a). The late Paleozoic Ancestral Rocky Mountains would have further complicated the courses of transcontinental river systems (Figs. 10E and 10F). During Paleozoic marine highstands, southern Appalachian rivers would have drained into midcontinent sedimentary basins and would not have reached the
Western Interior (Figs. 10B–10F). However, during the late Pennsylvanian and Permian, southern Appalachian rivers may have flowed directly into the northern Western Interior during marine lowstands. On Figures 10E and 10F, the areas across the midcontinent shown as epicontinental seas were likely exposed during lowstands from southern Canada to central Texas; this condition would have permitted direct fluvial or eolian transport from the southern Appalachians into the Dakotas, Wyoming, and possibly Montana.

These complications in Paleozoic paleogeography suggest that few, if any, Appalachian rivers directly reached the Grand Canyon region, and that wind systems may have played a significant role. During the middle and late Paleozoic, northeasterly trade winds (northerly in present coordinate systems) likely deflated Appalachian river systems that reached the northern United States and southern and central Canada, as described previously, and transported sand and loess southwestward. The trail of late Paleozoic eolian sandstones ranges from Lower Pennsylvanian (Atokan) in the Grand Canyon (Manacacha Formation) through Middle and Upper Pennsylvanian across much of the Western Interior (Hermosa, Weber, Casper, Wescogame, etc., Formations), through Permian (Weber, Cedar Mesa, Esplanade, etc., Formations). Consistent northeasterly winds transported sand southwestward throughout the late Paleozoic (Peterson, 1988).

Given that Grand Canyon strata contain abundant detrital zircons that are interpreted to have been shed from the southern and central Appalachian orogen, it is tempting to interpret the age distributions of these units as a possible record of magmatism in the Appalachian orogen. As shown in Figure 8D, periods of high magmatic flux appear to have occurred during 520–750 Ma (with peaks at 570 and 625 Ma), 380–490 Ma (with peaks at 424 Ma and possibly 384 and 412 Ma), and 268–365 Ma (with peaks at 310, 338, and 352 Ma). In comparison with the age distribution from Appalachian strata (Fig. 8D), high magmatic flux during Ordovician–Silurian (Taconic–early Acadian) time is apparent in both but in lower proportion in the Neoproterozoic magmatism during Appalachian strata (Fig. 8D), 338, and 352 Ma). In comparison with the age distribution from Appalachian strata (Fig. 8D), 338, and 352 Ma). In comparison with the age distribution from Appalachian strata (Fig. 8D), the crystalline terranes within portions of the U.S. Appalachian orogen, the crystalline terranes within portions of the U.S.

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