# Isotopic age of the Black Forest Bed, Petrified Forest Member, Chinle Formation, Arizona: An example of dating a continental sandstone

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## ABSTRACT

**Zircons from the Black Forest Bed, Petrified Forest Member, Chinle Formation, in Petrified Forest National Park, yield ages that range from Late Triassic to Late Archean. Grains were analyzed by multigrain TIMS (thermal-ionization mass spectrometry), single-crystal TIMS, and SHRIMP (sensitive, high-resolution ion-microprobe). Multiple-grain analysis yielded a discordia trajectory with a lower intercept of 207 ± 2 Ma, which because of the nature of multiple-grain sampling of a detrital bed, is not considered conclusive. Analysis of 29 detrital-zircon grains by TIMS yielded U-Pb ages of 2706 ± 6 Ma to 206 ± 6 Ma. Eleven of these ages lie between 211 and 216 ± 6.8 Ma. Our statistical analysis of these grains indicates that the mean of the ages, 213 ± 1.7 Ma, reflects more analytical error than geologic variability in sources of the grains. Grains with ages of ca. 1400 Ma were derived from the widespread plutons of that age exposed throughout the southwestern Cordillera and central United States. Twelve grains analyzed by SHRIMP provide $^{206}Pb^{*}/^{238}U$ ages from 214 ± 2 Ma to 200 ± 4 Ma. We use these data to infer that cores of inherited material were present in many zircons and that single-crystal analysis provides an accurate estimation of the age of the bed. We further propose that, even if some degree of reworking has occurred, the very strong concentration of ages at ca. 213 Ma provides a maximum age for the Black Forest Bed of 213 ± 1.7 Ma. The actual age of the bed may be closer to 209 Ma.**

**Dating continental successions is very difficult when distinct ash beds are not clearly identified, as is the case in the Chinle Formation. Detrital zircons in the Black Forest Bed, however, are dominated by an acicular morphology with preserved delicate terminations. The shape of these crystals and their inferred environment of deposition in slow-water settings suggest that the crystals were not far removed from their site of deposition in space and likely not far in time. Plinian ash clouds derived from explosive eruptions along the early Mesozoic Cordilleran margin provided the crystals to the Chinle basin, where local conditions insured their preservation. In the case of the Black Forest Bed, the products of one major eruption may dominate the volcanic contribution to the unit.**

**Volcanic detritus in the Chinle Formation was derived from multiple, distinct sources. Coarse pebble- to cobble-size material may have originated in eastern California and/or western Arizona, where Triassic plutons are exposed. Fine-grained detritus, in contrast, was carried in ash clouds that derived from caldera eruptions in east-central California or western Nevada.**

**Keywords: isotopic dating, Chinle Formation, Triassic, single-crystal TIMS, SHRIMP.**

## INTRODUCTION

Sedimentary deposits in continental settings are dated in most cases through fossil occurrences. These ages potentially provide worldwide stage correlations, but can only rarely be absolutely correlated with marine fauna, which commonly provide the initial stage reference. Isotopic dating, in contrast, provides a more certain reference frame for the time of deposition.

Isotopic dating of continental sequences is difficult in most circumstances. In arc terranes, where sedimentary rocks are interstratified with volcanic rocks, ages are easily obtained. Less commonly can these sedimentary rocks be correlated with backarc or purely continental units (e.g., Riggs et al., 1993a). Ash beds have been used with much success in this regard (e.g., Kowallis and Heaton, 1987; Kowallis et al., 1991, 2001; Christiansen et al., 1994; Blakey and Parnell, 1995), but are clearly confined to sequences that offer a high preservation potential for wind-carried material.

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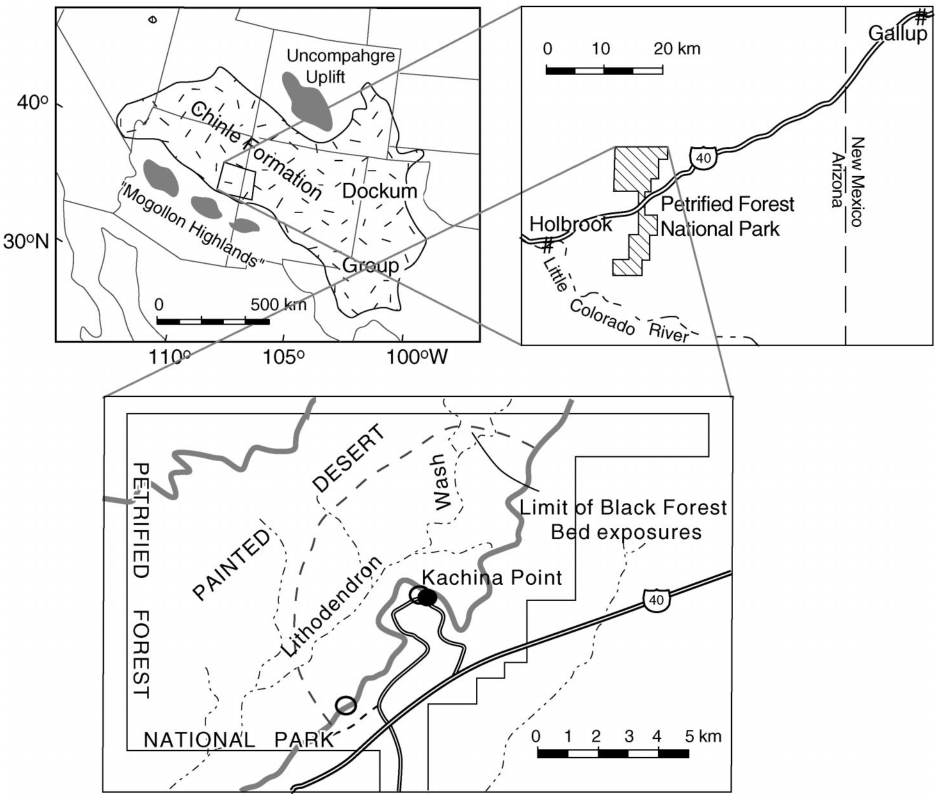


**Figure 1. Location map of Chinle Formation and Dockum Group (after Stewart et al., 1972), Petrified Forest National Park, and sample sites in the northern part of the park, with outcrop area of the Black Forest Bed enclosed by the dashed line. Closed circle—sample site for multigrain and single-grain samples; open circles—sample sites for multigrain fractions.**

The Late Triassic Chinle Formation is a continental sedimentary succession exposed on the Colorado Plateau; correlative units crop out from Texas to Nevada and from Arizona to Colorado (Stewart et al., 1972; Fig. 1). The Chinle Formation was deposited to the east and north of the incipient Triassic magmatic arc on the Cordilleran margin and contains abundant volcanic detritus. Fossil control suggests a Carnian–Norian age for the formation, and isotopic work has suggested Triassic to Jurassic ages (Peirce et al., 1985; Ash et al., 1986; Riggs et al., 1994a) for volcanic material.

Our sampling of an ash-rich sandstone from the Black Forest Bed of the Petrified Forest Member of the Chinle Formation has yielded zircon crystals that provide a maximum age of deposition of 213 ± 2 Ma, with a more likely age of deposition of ca. 209–202 Ma. This result represents the first significant dating of the Chinle Formation and demonstrates how geochronology can be applied to nonvolcanogenic sedimentary successions that lack distinct and coherent ash beds.

## CHINLE FORMATION

The Chinle Formation is as thick as ~600 m; it comprises several members that are exposed locally or are correlative across much of the outcrop area of the formation (Stewart et al., 1972; Dubiel, 1987, 1992). The Chinle Formation contains a rich array of Carnian–Norian floral and faunal fossil assemblages (Ash et al., 1986; Litwin et al., 1991). In addition, the presence of volcaniclastic detritus has been noted since early studies of the formation (Allen, 1930; Waters and Granger, 1953; Cadigan, 1963; Schultz, 1963; Stewart et al., 1972, 1986; Ash, 1989).

### Age of the Chinle Formation

The assigned age of the Chinle Formation is based primarily on the fossil accumulations (e.g., Fisher and Dunay, 1984; Ash et al., 1986; Litwin et al., 1991; Lucas et al., 1992). According to the time scale of Gradstein and Ogg (1996), the Carnian and Norian Stages ranged from 227 ± 5 to 221 ± 4 Ma and from 221 Ma to 210 ± 4 Ma, respectively. Rhaetian (latest Triassic: 210 Ma to 206 ± 4 Ma; Gradstein and Ogg, 1996) palynomorph assemblages are not recognized in the Chinle Formation (Litwin et al., 1991), but may be present (S. Lucas, 1994, personal commun.). Heckert and Lucas (1999) have suggested an essentially simultaneous worldwide first appearance of dinosaurs, including some found in the Chinle Formation, in late Carnian time.

Previous isotopic dating of the Chinle Formation has not been conclusive. Four dacite to andesite cobbles in the Petrified Forest Member yielded K-Ar ages of 222 ± 5 Ma to 196 ± 4 Ma (Peirce et al., 1985). Ash (1989, 1992) documented K-Ar analysis of biotite from the Black Forest Bed that yielded an age of 239 ± 9 Ma. Within error ranges, biotite analyzed by Ash (1989, 1992) and the older of ages of cobbles reported by Peirce et al. (1985) fall within the Carnian–Norian Stages. These reported ages, however, only provide potential maxima for the age of the formation itself. Peirce et al. (1985) acknowledged that the Petrified Forest Member is younger than the oldest of the cobbles that they analyzed, but that putting an undue emphasis on the younger cobble ages allows an age for the unit as young as ca. 195 Ma, contradicting a Norian age. Similarly, the biotite age reported by Ash (1989, 1992) is older than the early Norian age suggested by floral zonation (Ash et al., 1986). Ash (1992) pointed out that this biotite, if accurately dated, must have been reworked from older ash. Riggs et al. (1994a) reported an age of 207 ± 2 Ma for the Black Forest Bed, on the basis of analyses of multigrain zircon fractions.

### Source of Volcanic Detritus in the Chinle Formation

The source of the volcanic ash, crystals, and pebbles and cobbles in the Chinle Formation has been speculated upon since the earliest reports on the formation. The traditional assumption has been that all volcanic detritus was derived from similar, co-aerial sources. Paleocurrents in stream deposits throughout the formation indicate a source to the south for detritus, and a nearby highland source was proposed (e.g., Stewart et al., 1972). Stewart et al. (1986) later acknowledged that Mesozoic volcanic rocks to the south of the Colorado Plateau are all apparently younger than the Late Triassic age of the Chinle Formation and suggested that sources are either hidden beneath Cenozoic cover or were translated away from southern Arizona by a transcurrent tectonic feature. Stewart et al. (1986) also suggested that a far-traveled wind derivation was unlikely. Riggs et al. (1993b) proposed that the amount of volcanic material in the Chinle requires far too large of a volcanic field to be concealed beneath younger cover. They further suggested that fine-grained material was in fact far transported by wind, while agreeing that coarse-grained detritus was potentially derived from a removed southern source. The increasing awareness of numerous Triassic plutons in the Mojave Desert of western Arizona and eastern California (e.g., Barth et al., 1997), however, may indicate a source in that area from which the volcanic cover has been eroded in post-Triassic time. Reynolds et al. (1989) suggested that pre-Jurassic uplift in western Arizona may have eroded Triassic volcanoes, but Triassic plutonic rocks are not present in the area of uplift.

## AGE OF THE BLACK FOREST BED

The Black Forest Bed was formally designated and described by Ash (1992), following preliminary recognition by Cooley (1957, 1959) and Billingsley (1985). The Black Forest Bed is exposed in the northern part of Petrified Forest National Park (Fig. 1) and lies ~130 m above the Sonsela Sandstone Bed (Fig. 2) within the Upper Petrified Forest Member. The Black Forest Bed comprises limestone-nodule conglomerate and overlain by pink, white, and purple mudrock, shale, and sandstone (Ash, 1992) (Fig. 2). The thickness of the unit varies from a few centimeters to a maximum of 12.6 m (Ash, 1992).

Ash (1992) described the upper part of the Black Forest Bed as reworked tuff, on the basis of his interpretation of ''more than 50% pyroclastic tuff fragments'' (p. 63). Broken crystals suggestive of pyroclastic origin are common, and we argue that the zircon crystals were derived from a high-altitude Plinian cloud. It is more accurate, however, to call these strata tuffaceous sandstone and mudrock, because, overall, most of the material is quartz sand of a nonvolcanic origin (Shipman, 1999).

The Black Forest Bed was deposited in a fluvial environment that consisted in variable parts of stable and avulsing channels (Riggs et al., 1994b; Shipman et al., 1996; Shipman, 1999). Volcanic material is well preserved in crevasse-splay sand sheets and in the channel deposits that are elsewhere represented by lateral-accretion and channel-bar–migration deposits.

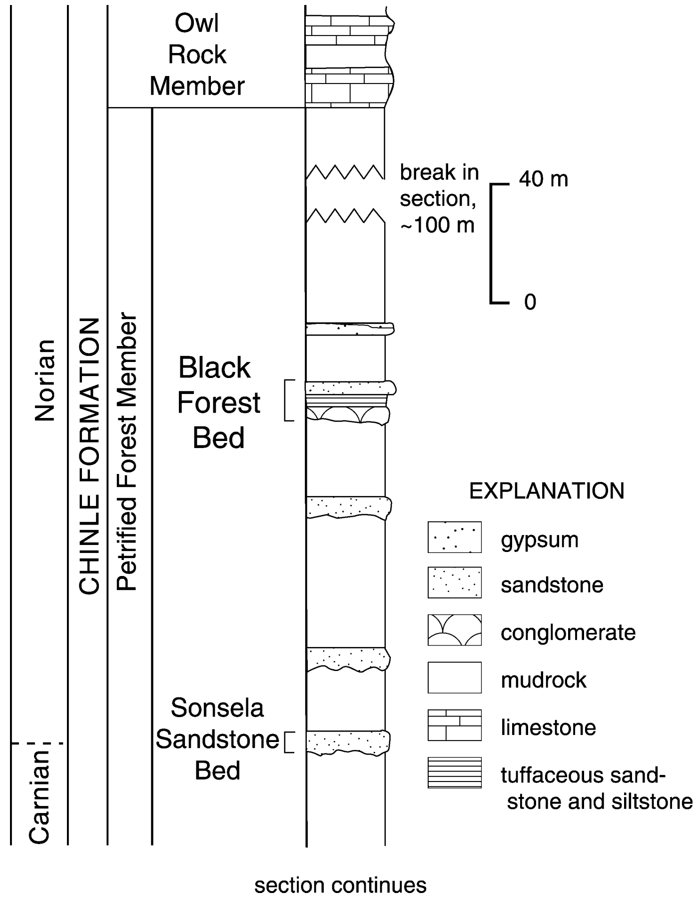


**Figure 2. Stratigraphic column of the Black Forest Bed, after Billingsley (1985); Ash (1992); and Dubiel et al. (1995)**

### Analytic Techniques

Zircons from the Black Forest Bed were analyzed both by multigrain TIMS and single-crystal TIMS and SHRIMP. Zircons analyzed by multigrain TIMS were sampled in three localities (Fig. 1), two within crevasse-splay deposits and one in a channel-bar deposit. The samples were crushed, and zircon crystals were extracted by standard techniques of heavy-mineral and magnetic separation. Separation on the basis of size, shape, and color yielded 12 multigrain fractions and 29 grains that were analyzed individually. For multigrain fractions, analytical methods followed conventional methods (modified from Krogh, 1973) using step-wise dissolution techniques described by Mattinson (1984, 1994). Errors were assigned following the approach of Mattinson (1987). Seventeen single grains were analyzed by TIMS, using methods described by Gehrels (2000). Twelve grains were analyzed on the SHRIMP-RG (sensitive high-resolution ion microprobe–reverse geometry) system at Stanford University. Individual zircons from conventional mineral separates were mounted in epoxy, polished, and coated with gold prior to analysis. Zircons were imaged with cathodoluminescence (CL) prior to analysis to guide selection of analysis points in zoned grains.

### Results

Fractions included stubby, acicular, and rounded shapes and smoky, pink, and clear colors. Of these, five multigrain fractions were clear and acicular and yielded two concordant ages (Table 1). Nonacicular zircons, or those with inclusions within the zircon crystals, yielded only discordant data. In this discussion, errors are stated as $2\sigma$.

When plotted on a Wetherill-type concordia diagram (Fig. 3) through the use of Ludwig's (1994) regression program, the five acicular-zircon fractions plot on a discordia line with a lower intercept of 207 $\pm$ 1.5 Ma (MSWD [mean square of weighted deviates] = 2.7). Clearly, some of the fractions contained zircons with a Proterozoic inherited component,

TABLE 1. ISOTOPIC DATA FROM TIMS ANALYSES OF MULTIGRAIN FRACTIONS

| Grain type† | Sample mass (mg) | $^{206}$Pb* (ppm) | $^{238}$U (ppm) | Isotopic ratios | | | | | | Ages | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | $^{204}$Pb/$^{206}$Pb | $^{208}$Pb/$^{206}$Pb | $^{207}$Pb/$^{206}$Pb | $^{206}$Pb*/$^{238}$U | $^{207}$Pb*/$^{235}$U | $^{207}$Pb*/$^{206}$Pb* | $^{206}$Pb*/$^{238}$U | $^{207}$Pb/$^{235}$U | $^{207}$Pb*/$^{206}$Pb* |
| PF-1 | | | | | | | | | | | | |
| A, d | 1 | 1.3 | 13.7 | 0.001638 | 0.28019 | 0.10850 | 0.10918 | 1.29353 | 0.08593 | 668.0 | 842.9 | 1337 ± 5 |
| A, e | 0.7 | 2.22 | 78 | 0.001276 | 0.30412 | 0.06882 | 0.03284 | 0.22890 | 0.05055 | 208.3 | 209.3 | 220 ± 6 |
| B, e | 3.2 | 4.4 | 141.7 | 0.000682 | 0.29273 | 0.06467 | 0.03590 | 0.27152 | 0.05485 | 227.4 | 243.9 | 406 ± 3 |
| B, d | 1.7 | 8.51 | 143.4 | 0.000212 | 0.18484 | 0.08217 | 0.06859 | 0.74997 | 0.07931 | 427.6 | 568.2 | 1180 ± 2 |
| B, g | 1.2 | 15.42 | 119.9 | 0.000212 | 0.15340 | 0.10126 | 0.14870 | 2.01922 | 0.09849 | 893.7 | 1122.0 | 1596 ± 2 |
| C, e | 2 | 2.26 | 73.7 | 0.000796 | 0.30776 | 0.06564 | 0.03548 | 0.26511 | 0.05419 | 224.8 | 238.8 | 379 ± 4 |
| PF-2 | | | | | | | | | | | | |
| A, e | 1.4 | 1.04 | 36.8 | 0.001158 | 0.22676 | 0.06692 | 0.03259 | 0.22632 | 0.05037 | 206.7 | 207.2 | 212 ± 6 |
| A, f | 2 | 1.22 | 39.3 | 0.000743 | 0.22998 | 0.06477 | 0.03595 | 0.26840 | 0.05415 | 227.7 | 241.4 | 377 ± 3 |
| B, e | 1 | 4.62 | 157.7 | 0.000524 | 0.26766 | 0.05973 | 0.03383 | 0.24356 | 0.05222 | 214.5 | 221.3 | 295 ± 3 |

*Notes*: Analyses done by N.R. Riggs at University of California, Santa Barbara. All samples nonmagnetic at 0.5° side tilt, 15° forward tilt on Frantz isodynamic separator. $^{206}$Pb/$^{204}$Pb is measured ratio, uncorrected for blank, spike, or fractionation. All uncertainties are at the 95% confidence level (2σ). Uncertainties in ages are given in millions of years. Constants used: $^{238}$U/$^{235}$U = 137.88; decay constant for $^{235}$U = $9.8485 \times 10^{-10}$; decay constant for $^{238}$U = $1.55125 \times 10^{-10}$. Mass-dependent corrections factors of 0.125% per mass unit. Uncertainties in $^{208}$Pb/$^{206}$Pb and $^{207}$Pb/$^{206}$Pb are <0.9%; uncertainties in $^{204}$Pb/$^{206}$Pb vary from 0.08% to 0.25%. Pb ratios corrected for 0.010 ng blank; U has been adjusted for 0.001 ng blank. Initial Pb from Stacey and Kramers (1975) using $^{208}$Pb:$^{207}$Pb:$^{206}$Pb:$^{204}$Pb of 38.5:15.6:18.4:1. Pb* denotes radiogenic Pb.

†Sizes and shapes of zircon crystals: A—100–150 m; B—80–100 μm; C—45–100 μm; d—clear, stubby; e—acicular; f—amber-color, stubby; g—clear, subround.

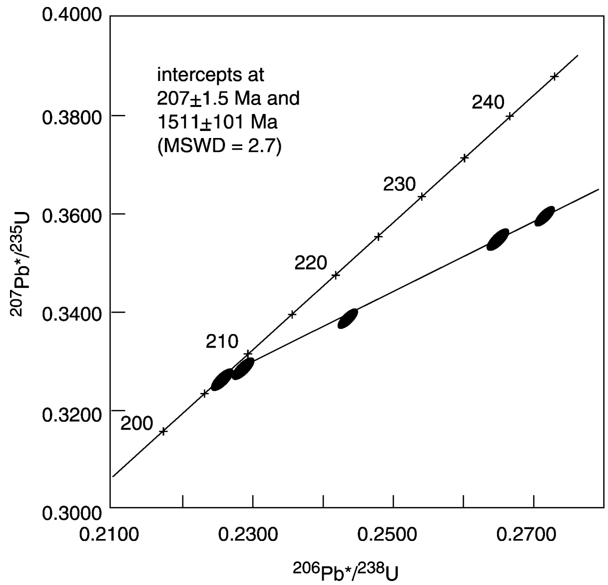


**Figure 3. Wetherill concordia diagram for multigrain fractions.**

but the suggested age of inheritance, 1511 ± 101 Ma, is geologically very reasonable for mixed ca. 1430 Ma and ca. 1700–1600 Ma grains derived from Proterozoic continental crust in western North America.

Single crystals comprised the same colors and shapes as multigrain fractions, including acicular zircons. Grains analyzed by TIMS range in $^{206}$Pb*/$^{238}$U age from 2706 ± 6 Ma to 211 ± 3.8 Ma (Fig. 4A, Table 2). The oldest grains may represent a recycled Gondwanan source, and several grains with ages between 1442 ± 9 Ma and 1406 ± 16 Ma ($^{207}$Pb*/$^{206}$Pb*) probably reflect 1500–1300 Ma granite sources throughout the North American Southwest (Anderson, 1989; Van Schmus et al., 1993).

Twelve crystals range in $^{206}$Pb*/$^{238}$U age between 224 ± 3.3 and 211 ± 3.8 Ma (Table 2). Of these, 11 are between 216 ± 6.8 and 211 Ma (although this point overlaps 224 ± 3.3 Ma, within errors, whereas 211 ± 3.8 does not) (Fig. 4B).

Grains analyzed by SHRIMP yielded $^{206}$Pb*/$^{238}$U ages between 214 ± 2 Ma and 200 ± 4 Ma (Table 3, Fig. 5). The oldest and youngest ages are discordant of one another on the basis of the errors on those ages (Fig. 5). Cathodoluminescence of the crystals was used largely to assess the likelihood of cores or zones that might yield inaccurate older ages in single crystals. The CL images suggest that the ages do not correlate with shape of zircon, whether described as degree of rounding, length, width, or aspect ratio. Dark, round to ragged areas within some crystals, however, are suggestive of the presence of cores, but the SHRIMP beam diameter was too large to uniquely sample any of these areas, and they were intentionally avoided when positioning the beam.

## Interpretation of Mesozoic Ages

Detrital-zircon age analysis is commonly used to study provenance and tectonic problems in sedimentary units (e.g., Gehrels and Dickinson, 1995; Riggs et al., 1996; Ross et al., 2001; Stewart et al., 2001; Gleason et al., 2002). In contrast, we propose that the dates reported herein can, at a minimum, provide an accurate maximum age of a continental bed.

### *Multigrain TIMS*

We report the multigrain fractions in the interests of completeness, but, in contrast to the conclusion of Riggs et al. (1994a), we think that the presence of numerous grains in a fraction gives rise to the strong possibility of mixing. Thus, we do not consider these data trustworthy in uniquely interpreting the age of the Black Forest Bed.

### *Single-Crystal TIMS*

The vast majority of zircons in the sample lie between the $^{206}$Pb*/$^{238}$U ages of 216 ± 6.8 and 211 ± 3.8 Ma. Although this age range reflects in part, a sampling bias (i.e., preferential picking of acicular zircons), all acicular zircons except one (together with four of other morphologies) lie within this range (Table 2). For this reason, we performed a statistical analysis to assess whether the close clustering of ages could be considered significant.

In the first case, we looked at a simple statistical analysis of the 11 zircons with $^{206}$Pb*/$^{238}$U ages between 216 and 211 Ma. The mean

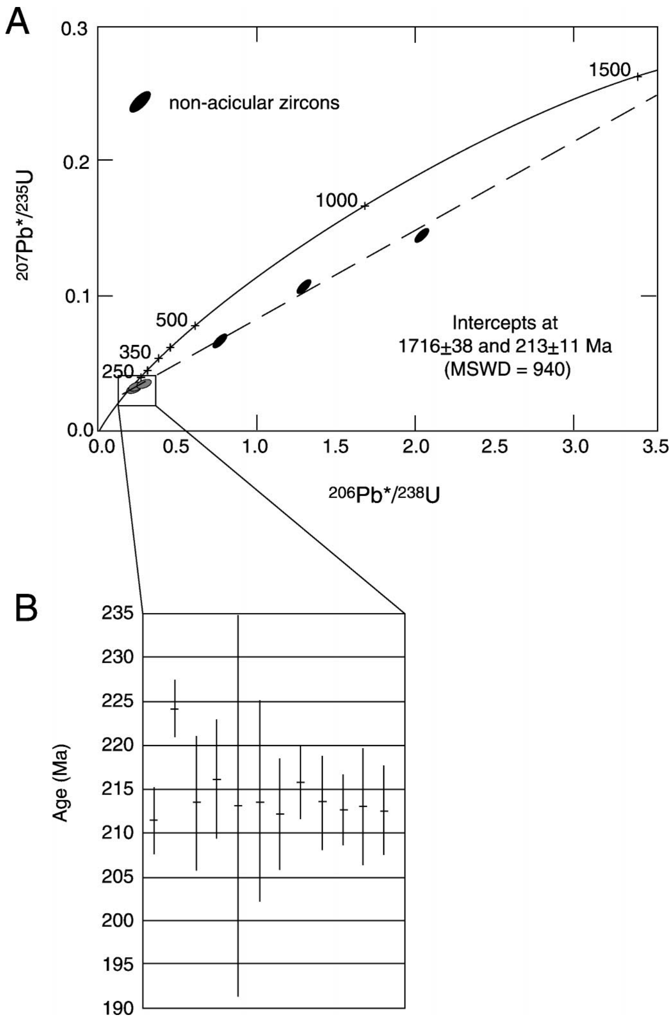


**Figure 4. (A) Wetherill concordia diagram for single-crystal TIMS analyses. (B) Simplified plot showing distribution of Mesozoic zircon ages.**

$^{206}$Pb*/$^{238}$U age of the zircons is 213 Ma, with a 2σ standard error of 1.7 Ma (1σ error = 0.83 Ma), which assumes that the grains are comagmatic and that the error is limited to parameters internal to the zircons and to the analysis (i.e., not a variation in sources). Because the crystals are detrital, which implies a geologic error superposed on the analytical error, we then applied an analysis in which $\sigma_a/n^{1/2}$ [the standard deviation of all the ages ($\sigma_a$) divided by the square root of the number of data points ($n^{1/2}$)] was compared to the 1σ analytical error derived from standard regression. For these samples, $\sigma_a/n^{1/2}$ is 1.01 m.y., which is close enough to the 0.83 m.y. 1σ standard error to imply very little superposed geologic variability (Table 2).

In the second case, we analyzed whether the ages are normally distributed about the mean, with errors entirely accounted for by measurement errors. This assessment involves solving the equation $x = m/\sigma_{ae}$, where $\sigma_{ae}$ is the 1σ analytical error, for each sample and $m$ = the mean age of that sample. If 63% of the data (i.e., 63% of the $x$ values) are between 1 and −1, the analytical errors account for a statistically relevant part of the discrepancies between samples. As shown in Table 2, only one analysis lies outside of the range +1 to −1, and therefore, 91% concordance is demonstrated. Thus, statistically, the analytical errors explain the variation in the data, without evidence of geologic variability like variable sources. We use the single-crystal TIMS data to interpret the age of the Black Forest bed to be 213 ± 1.7 Ma.

### *Single-Crystal SHRIMP*

Although $^{206}$Pb*/$^{238}$U ages have relatively small associated errors, when the statistical test just described is applied to the SHRIMP data, a mean age of 209 ± 5 Ma is obtained, but the sample is only 42% concordant. This result strongly suggests geologic variability in the [SHRIMP] sample population. For these reasons, we propose that the maximum age of the Black Forest Bed is ca. 214 Ma, in agreement with the single-crystal TIMS data, and that a more likely age is between 209 Ma and 202 Ma, which reflects the average of the several youngest spot ages and the mean age of the entire sample.

## Conflicts with Previously Obtained Ages and Potential Sources of Error

A maximum isotopic age for the upper Petrified Forest Member of 214 Ma is younger than most of the previous estimates of the age of the formation. In addition, it is younger than other strata that contain dinosaur fauna similar to those in the Chinle Formation (Lucas et al., 1992; Rogers et al., 1993). These discrepancies suggest a potential error in the analyses, and two obvious possibilities must be considered.

The Chinle Formation is overlain by the Pleistocene Bidahochi Formation, which includes a series of lava flows and small, 7 Ma maar and cinder volcanoes (Ort et al., 1998). A maar volcano is reported ∼1 km from the sample sites, suggesting that intrusive phases related to Bidahochi Formation volcanism may have intruded through the Black Forest Bed. Any heat from a dike or lava flow, however, would be dissipated in far less than the 1 km lateral distance between the sample sites and the volcano or the ∼50 m of elevation between the lava flow and the strata. Thus, we strongly doubt that Pleistocene localized heating caused disruption of the isotopic systems in the zircon crystals.

The Black Forest Bed represents a mixture of several different provenances. As such, any particular grain population, whether quartz, feldspar, or zircon, represents a mixture from any of several sources. Zircon that yielded a 224 ± 3.3 Ma $^{206}$Pb*/$^{238}$U age may indicate that some material was reworked from a different source, but equally may represent inheritance in a 213 Ma or younger grain. Likewise, the 200 ± 5 Ma zircon may indicate that the unit is younger than proposed here. The statistical tightness of the single-crystal TIMS data suggests that the mean age, 213 ± 1.5 Ma, is geologically significant. The presence of likely cores, however, suggests that many of the crystals have an older component to them, however small, potentially making the 213 Ma age slightly older than the actual age of the eruption that supplied the majority of the crystals. The age of the cores could be as

TABLE 2. ISOTOPIC DATA FROM TIMS ANALYSES OF SINGLE ZIRCON CRYSTALS

| Grain type† | Grain mass (μg) | $Pb_c$ (pg) | U (ppm) | Isotopic ratios | | | | Ages | | | m/1σ error |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | $^{206}Pb_m/^{204}Pb$ | $^{206}Pb_c/^{208}Pb$ | $^{206}Pb^*/^{238}U$ | $^{207}Pb^*/^{235}U$ | $^{206}Pb^*/^{238}U$ | $^{207}Pb^*/^{235}U$ | $^{207}Pb^*/^{206}Pb^*$ | |
| C1 | 8 | 11 | 38 | 869 | 2.7 | 0.51887 ± 1.44 | 13.2964 ± 1.49 | 2694 | 2701 | 2706 ± 6 | |
| A1 | 11 | 11 | 74 | 1034 | 3.9 | 0.21922 ± 1.05 | 2.6921 ± 1.39 | 1278 | 1326 | 1406 ± 16 | |
| A2 | 18 | 15 | 149 | 2412 | 6.0 | 0.22388 ± 0.65 | 2.7780 ± 0.81 | 1302 | 1350 | 1425 ± 9 | |
| A3 | 15 | 16 | 188 | 2339 | 8.3 | 0.22000 ± 0.70 | 2.7539 ± 0.86 | 1282 | 1343 | 1442 ± 9 | |
| B1 | 15 | 14 | 165 | 379 | 4.0 | 0.03333 ± 1.80 | 0.2301 ± 3.42 | 211 ± 4 | 210 | 198 ± 64 | −.99 |
| B2 | 9 | 18 | 366 | 416 | 3.5 | 0.03538 ± 1.46 | 0.2445 ± 3.31 | 224 ± 3 | 222 | 200 ± 66 | |
| B3 | 12 | 20 | 93 | 138 | 2.1 | 0.03367 ± 3.61 | 0.2319 ± 7.91 | 213 ± 8 | 212 | 193 ± 160 | 0.06 |
| D1 | 11 | 22 | 116 | 141 | 1.2 | 0.03411 ± 3.17 | 0.2414 ± 7.10 | 216 ± 7 | 220 | 256 ± 140 | 0.86 |
| D2 | 7 | 11 | 55 | 97 | 1.6 | 0.03360 ± 10.20 | 0.2301 ± 14.00 | 213 ± 22 | 210 | 180 ± 220 | −0.02 |
| E1 | 18 | 12 | 151 | 2989 | 5.1 | 0.21156 ± 1.01 | 2.6161 ± 1.08 | 1237 | 1305 | 1419 ± 7 | |
| E2 | 14 | 8 | 57 | 234 | 2.2 | 0.03369 ± 5.43 | 0.2421 ± 6.84 | 214 ± 12 | 220 | 290 ± 93 | 0.06 |
| E3 | 18 | 15 | 84 | 229 | 3.2 | 0.03346 ± 3.02 | 0.2316 ± 5.33 | 212 ± 6 | 212 | 205 ± 99 | −0.34 |
| E4 | 15 | 16 | 185 | 378 | 3.8 | 0.03404 ± 1.95 | 0.2358 ± 3.85 | 216 ± 4 | 215 | 206 ± 74 | 1.20 |
| E5 | 12 | 9 | 148 | 419 | 5.4 | 0.03367 ± 2.49 | 0.2294 ± 4.38 | 213 ± 5 | 210 | 168 ± 82 | 0.08 |
| E6 | 18 | 12 | 125 | 404 | 3.8 | 0.03354 ± 1.95 | 0.2307 ± 3.43 | 213 ± 4 | 211 | 190 ± 63 | −0.03 |
| E7 | 15 | 23 | 88 | 135 | 1.6 | 0.03358 ± 3.13 | 0.2326 ± 6.91 | 213 ± 7 | 212 | 206 ± 140 | −0.10 |
| E8 | 19 | 29 | 92 | 145 | 2.0 | 0.03352 ± 2.38 | 0.2347 ± 5.36 | 213 ± 5 | 214 | 231 ± 110 | −0.27 |

*Notes*: Analyses done by G.E. Gehrels, University of Arizona Laboratory of Geochronology. All grains abraded in air-abrasion device and analyzed as single crystals. $^{206}Pb_m/^{204}Pb$ is measured ratio, uncorrected for blank, spike, or fractionation. $^{206}Pb_c/^{208}Pb$ is corrected for blank, spike, and fractionation. All uncertainties are at the 95% confidence level (2σ). Uncertainties in isotope ratios are in percent. Uncertainties in ages are in millions of years. Most concentrations have an uncertainty of 25% owing to uncertainty in mass of grain. Constants used: $^{238}U/^{235}U = 137.88$; decay constant for $^{235}U = 9.8485 \times 10^{-10}$; decay constant for $^{238}U = 1.55125 \times 10^{-10}$. Isotope ratios are adjusted as follows: (1) Mass-dependent corrections factors of: 0.14 ± 0.06 %/amu for Pb and 0.04 ± 0.04 %/amu for $UO_2$. (2) Pb ratios corrected for 0.005 ± 0.003 ng blank with $^{206}Pb/^{204}Pb = 18.6 \pm 0.3$, $^{207}Pb/^{204}Pb = 15.5 \pm 0.3$, and $^{208}Pb/^{204}Pb = 38.0 \pm 0.8$. (3) U has been adjusted for 0.001 ± 0.001 ng blank. (4) Initial Pb from Stacey and Kramers (1975), with uncertainties of 1.0 for $^{206}Pb/^{204}Pb$, 0.3 for $^{207}Pb/^{204}Pb$, and 2.0 for $^{208}Pb/^{204}Pb$. Pb* denotes radiogenic Pb.

†Grain shapes: A—barrel-shaped, pink to purple; B—subhedral, clear to yellow; aspect ratio 2–3:1; C—well-rounded, purple; D—sub- to euhedral, clear, aspect ratio 3–4:1; E—clear, acicular.

TABLE 3. ISOTOPIC DATA FROM SHRIMP-RG ANALYSES OF SINGLE CRYSTALS

| Spot | U (ppm) | Th (ppm) | Isotopic ratios | | | | | | Age $^{206}Pb^*/^{238}U$ |
|---|---|---|---|---|---|---|---|---|---|
| | | | $^{206}Pb/^{238}U$ | Error | $^{207}Pb/^{206}Pb$ | Error | $^{206}Pb^*/^{238}U$ | Error | |
| PF12-1 | 77 | 67 | 0.02226 | 7.0 | 0.05033 | 16.6 | 0.03176 | 3.7 | 202 ± 7 |
| PF12-2 | 215 | 233 | 0.02113 | 11.5 | 0.05147 | 9.2 | 0.03362 | 2.1 | 213 ± 4 |
| PF12-3 | 192 | 174 | 0.02410 | 8.4 | 0.05082 | 10.0 | 0.03371 | 2.2 | 214 ± 5 |
| PF12-4 | 540 | 399 | 0.02388 | 4.2 | 0.05024 | 5.8 | 0.03340 | 1.3 | 212 ± 3 |
| PF12-5 | 120 | 178 | 0.02484 | 7.5 | 0.05448 | 14.4 | 0.03255 | 3.4 | 207 ± 7 |
| PF12-6 | 487 | 417 | 0.02108 | 3.4 | 0.04807 | 7.0 | 0.03341 | 1.7 | 212 ± 4 |
| PF12-7 | 411 | 138 | 0.02156 | 5.6 | 0.05055 | 7.4 | 0.03235 | 1.6 | 205 ± 3 |
| PF12-8 | 124 | 119 | 0.02406 | 7.7 | 0.05695 | 12.4 | 0.03259 | 3.2 | 207 ± 7 |
| PF12-9 | 332 | 686 | 0.02523 | 2.7 | 0.05226 | 7.8 | 0.03204 | 1.9 | 203 ± 4 |
| PF12-10 | 179 | 151 | 0.02575 | 4.1 | 0.05197 | 11.4 | 0.03158 | 2.9 | 200 ± 6 |
| PF12-11 | 305 | 147 | 0.02794 | 2.5 | 0.04971 | 8.4 | 0.03306 | 3.3 | 210 ± 7 |
| PF12-12 | 127 | 145 | 0.02419 | 3.9 | 0.05052 | 12.6 | 0.03248 | 3.1 | 206 ± 6 |

*Notes*: Analyses done by J.L. Wooden, USGS-SUMAC, Stanford University. Isotopic ratio errors in percent, 2σ. Data are corrected to an age of 419 Ma (R. Mundil, 1999, and S.L. Kamo, 2001**, personal communications to J. Aleinikoff; conventional TIMS ages) for standard R33. Errors on $^{206}Pb^*/^{238}U$ ages in m.y., 2σ; ratios and errors corrected for excess $^{207}Pb$. Pb* denotes radiogenic Pb.

similar as 5–10 m.y. older than the main body of the zircon crystal or as dissimilar as Proterozoic. The narrow clustering of ages leads us to speculate that the former is more likely.

Analysis of biotite by $^{40}Ar/^{39}Ar$ methods and a search for sanidine crystals in the sandstones yielded highly uncertain and inconclusive results. Apparently, sanidines are not present in the sandstone, and biotites have a highly diverse origin.

## DISCUSSION

Faunal data specific to the Black Forest Bed and the Petrified Forest Member are based on palynomorph assemblages. Palynomorph assemblages in the Petrified Forest Member stratigraphically near the Sonsela Sandstone Bed indicate a ''younger than late Carnian, but older than the latest Norian'' age (Litwin et al., 1991, p. 280).

Dating of a tuff horizon in Argentina has led to assignment of an age of 227.8 ± 0.3 Ma for the Carnian Ischigualasto tetrapod assemblage (Rogers et al., 1993), including fauna similar to those found in the Chinle Formation (Lucas et al., 1992). These similar tetrapod fauna in the Chinle Formation are found in the lower Petrified Forest Member (R.F. Dubiel, 1994, personal commun.; Blue Mesa Sandstone of Lucas et al., 1992), ~120 m down section from the Black Forest Bed. This fact implies either a 15–20 m.y. time span to deposit 120 m of the Chinle Formation, including periods of slow deposition and erosion and/or unconformities, or a lack of synchronicity in appearance of the dinosaur fauna.

## Source of Zircons in the Black Forest Bed

Stewart et al. (1986) suggested that a far-traveled source for volcanic detritus in the Chinle Formation was unlikely, in large part on the basis of the size of abundant volcanic cobbles and pebbles. In contrast, sand-grained-sized material is more likely to have come from a great distance. Plinian clouds derived from large, explosive volcanic eruptions commonly carry ash and crystals for hundreds of kilometers at tropospheric and stratospheric levels. For example, a 0.3-m-thick ash bed correlative with the prehistoric eruption of the Valles, New Mexico, caldera has been identified in Texas, over 500 km from the source volcano (Izett et al., 1972). Deposition from these clouds yields ash beds in some cases (e.g., Kowallis and Heaton, 1987; Kowallis et al., 1991) and ash-rich sediment in others (e.g., much of the Chinle Formation).

The ranges of zircon grain sizes, the acicular shape of the zircons, and the environment of deposition suggest that crystals did not travel great distances once deposited from an ash cloud. A very fine quartz grain (the average grain size of the Black Forest Bed) can be buffered in the suspended load of a stream for long distances, but the larger acicular zircons (~100 μm), being denser than quartz,

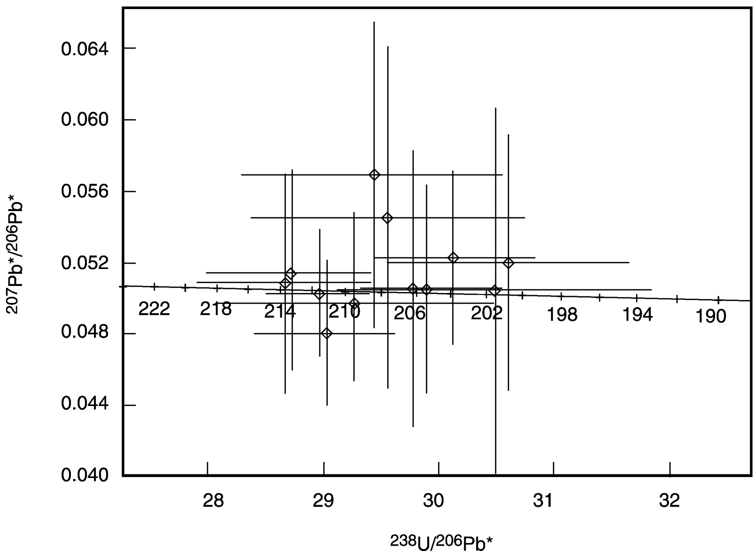


**Figure 5. Tera-Wasserburg diagram showing age distribution (in Ma) of grains analyzed by SHRIMP.**

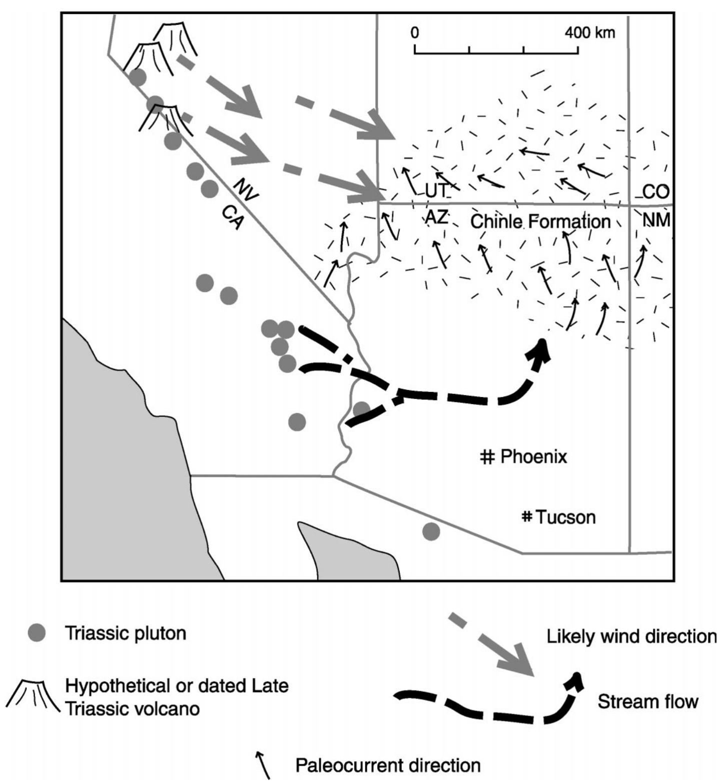


**Figure 6. Map showing speculative sources of Chinle volcanic detritus. Fine-grained airborne detritus was possibly derived from volcanoes in eastern California and/or western Nevada; coarse-grained detritus was carried in fluvial systems, possibly from sources in westernmost Arizona and southeastern California (Triassic plutons and volcanoes after Snow et al., 1991; Schweickert and Lahren, 1993; Miller et al., 1995; Barth et al., 1997). Paleocurrents after Stewart et al. (1972). Plutons (but not state boundaries!) palinspastically reconstructed ~85 km northeast of their present locations (after Spencer and Reynolds, 1991); reconstruction does not include shortening in the Maria fold belt.**

would probably be hydrologically unstable, and their delicate form would not likely be preserved. The settling velocity for a large zircon is equivalent to that of a fine to medium quartz grain, a size at which intermittent suspension is common (Boggs, 1987) and abrasion could occur. The preservation of these grains indicates a minimum of transport, in turn suggesting derivation from an ash cloud and deposition in a relatively quiet environment.

The zircons, derived from a Plinian ash cloud, either may have been deposited by fallout into the quiet waters of an overbank deposit or, soon after deposition in a stream, may have been washed into crevasse-splay sheets (Shipman, 1999). The range in settling velocities of the largest to smallest zircon and quartz grains suggests either an abrupt change in stream energy or relatively unsorted deposition into the quiescent environment. In either case, a minimum of transport is required before the grains were dropped from suspension. Also in either case, the age of the zircons represents a very near approximation of the age of the Black Forest Bed.

It can be effectively argued that older zircon-bearing pyroclastic units at the source volcanoes were eroded into a stream and that the date given by the zircons represents a maximum, but not a precise, age for the deposit. For this reason we propose that the age of the Black Forest Bed is 209 $\pm$ 5 Ma, nearly certainly younger than 214 Ma and potentially as young as 200 $\pm$ 5 Ma.

**Location of Source Volcanoes**

Ash deposits from Plinian clouds very commonly reflect the dominant wind direction across the volcano and regionally across a land mass (e.g., Izett et al., 1972; Sarna-Wojcicki et al., 1981). Ash beds are also found in directions contrary to apparent primary wind direction from the eruptive center; for example, ash beds from the eruption of Long Valley caldera in southeastern California have been found in central and coastal California and in Deep Sea Drilling Project cores (Sarna-Wojcicki et al., 1984, 1987). In the case of the Chinle Formation, a lack of paleowind direction indicators compounds the problem of considering the direction of source volcanoes.

A compilation of cross-bedding data by Peterson (1988) suggests that wind directions did not change substantially from Pennsylvanian to Middle Jurassic time. Although paleowind data are not available from Upper Triassic deposits, it may be reasonable, although certainly not required, to infer a wind direction from the north or northwest, as inferred by Peterson (1988).

Large volcanic clasts in the Chinle Formation must have had an element of derivation from the south, as indicated by paleocurrents (Stewart et al., 1972). Volcanic rocks of certain Triassic age are missing from areas south, southeast, and southwest of the Chinle Formation (Stewart et al., 1986), but plutonic bodies throughout the Mojave Desert of eastern California and western Arizona yield Triassic ages (Snow et al., 1991; Miller et al., 1995; Barth et al., 1997). These plutons may be the roots of a volcanic arc that certainly could have fed volcanic flows large enough to contribute the detritus seen in the lower Chinle Formation. Volcanic clasts are rare in outcrops of Chinle Formation in southern Nevada, however, suggesting that streams draining volcanic highlands in southeastern California or western Arizona flowed to the east before turning north (Fig. 6).

Silicic igneous rocks of Triassic age, including volcanic rocks and plutonic rocks that may be the roots of volcanoes, are exposed in west-central Nevada and in pendants in the central Sierra Nevada, and their ages include a range from 215 to 205 Ma (Dilles and Wright, 1988; Schweickert and Lahren, 1989; Schweickert and Lahren, 1993). Caldera complexes in the southern Sierra Nevada are marine (Busby-Spera, 1986; C.J. Busby, 2002, personal commun.) but may have contributed to a large Plinian cloud. The presence of volcanic rocks and their roots in western Nevada, in east-central and southeastern California, and in western Arizona provides a reasonable source for ash and crystals and potentially for coarse detritus in the Chinle Formation (Fig. 6).

Lastly, Stewart et al. (1997 and 2002, personal commun.) have proposed that plutons currently exposed in western Nevada and eastern California were moved along transcurrent faults from original positions in northern Sonora. Although more difficult to evaluate, this model provides a prospective source of detritus to the south of the Chinle basin, allowing paleocurrents to reflect original source directions.

## CONCLUSIONS

Detrital acicular zircon crystals in the Black Forest Bed, Petrified Forest Member, Chinle Formation, yield ages that indicate a likely depositional age of 209 $\pm$ 5 Ma for deposition of the bed; a maximum is certainly given by 213 $\pm$ 1.7 Ma. Detrital minerals in a continental unit that lacks distinct ash beds are rarely considered to date the actual deposition of the bed. We suggest, however, that the delicate morphology of the zircons and their deposition in crevasse-splay and overbank settings indicate that they were not far travelled, either in time or space. The tight clustering of their ages likely provides a very near age for this succession. Zircons may have been derived dominantly from one large explosive volcanic eruption. The source of the coarser volcanic detritus remains obscure, but may have been the volcanic rocks once present in the Mojave Desert or the Triassic volcanoes of west-central Nevada and the Sierra Nevada.

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