Paleogeographic reconstruction of the Eocene Idaho River, North American Cordillera

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ABSTRACT

Eocene Lake Gosuie in southwestern Wyoming was progressively filled in by volcaniclastic sediment between 49.6 and 47.0 Ma. The source of this material has long been thought to have been the Absaroka volcanic province, immediately north of the greater Green River Basin. Lead isotope compositions of sandstone from this interval, however, are consistent with derivation from the Challis volcanic field. The 40Ar/39Ar eruptive ages of single detrital K-feldspar crystals from greater Green River Basin sandstones are nearly identical to 40Ar/39Ar eruptive ages for volcanic rocks of the Challis volcanic field (49.8–45.5 Ma), but we also identify Mesozoic and Proterozoic detrital and other isotopic provenance techniques that provide a consistent record of the Challis volcanic field (49.8–45.5 Ma), but we also identify Mesozoic and Proterozoic detrital and other isotopic provenance techniques that provide a consistent record of the Challis volcanic field. The 40Ar/39Ar ages of single detrital K-feldspar crystals from the Challis volcanic field (49.8–45.5 Ma), but we also identify Mesozoic and Proterozoic detrital and other isotopic provenance techniques that provide a consistent record of the Challis volcanic field. The 40Ar/39Ar ages of single detrital K-feldspar crystals from the Challis volcanic field (49.8–45.5 Ma), but we also identify Mesozoic and Proterozoic detrital and other isotopic provenance techniques that provide a consistent record of the Challis volcanic field. The 40Ar/39Ar ages of single detrital K-feldspar crystals from the Challis volcanic field (49.8–45.5 Ma), but we also identify Mesozoic and Proterozoic detrital and other isotopic provenance techniques that provide a consistent record of the Challis volcanic field. The 40Ar/39Ar ages of single detrital K-feldspar crystals from the Challis volcanic field (49.8–45.5 Ma), but we also identify Mesozoic and Proterozoic detrital and other isotopic provenance techniques that provide a consistent record of the Challis volcanic field.

INTRODUCTION

Lacustrine sedimentary basins have long been exploited as important archives of paleoclimate due to their well-known ability to resolve highly detailed and localized histories of past climate change on the continents. Additionally, many contain important records of vertebrate (including hominid) evolution, and substantial economic resources (most notably evaporite minerals, coal, oil, and oil shale). The Eocene Green River Formation in the western United States is arguably the world’s most famous and best-documented assemblage of pre-Quaternary lake strata, as evidenced by ~2000 published books, papers, and abstracts dating back to its original naming by Hayden (1869). It is also a classic locality for interpreting Milankovitch-scale climatic forcing of sedimentation (e.g., Bradley, 1929; Fischer and Roberts, 1991; Roehler, 1993; Machlus et al., 2008; Meyers, 2008; Smith et al., 2010). Deposition of the Green River Formation also coincided with the most recent period of unusually warm climate (the early Eocene climatic optimum; Zachos et al., 2001; Smith et al., 2003), and thus it provides an unparalleled opportunity to examine potential feedbacks between warm climate and continental weathering (Smith et al., 2008a). Finally, the Green River Formation contains the world’s largest commercial deposits of both soda ash (Dy ni, 1996) and oil shale (Dy ni, 2006). Studies of ancient lake systems often fail to adequately consider modern analogs. Many large modern lakes receive drainage from regional or even interregional rivers, but ancient lake systems are often inferred to have received runoff only from relatively local watersheds (e.g., Dickinson et al., 1988; Smith et al., 2008b). In some cases, the source or significance of surface runoff is not considered at all. This difficulty commonly arises from the lack of adequate data to reconstruct the detailed paleohydrologic relationships between ancient rivers and the lakes they fed. Recent advances in detrital and other isotopic provenance techniques offer the opportunity to greatly ameliorate the problem (e.g., Heller et al., 1985; Riggs et al., 1996; DeGraaff-Surpless et al., 2002; Dickinson and Gehrels, 2008), but relatively few studies of ancient lake systems have yet adopted this approach (e.g., Rhodes et al., 2002; Gierlowski-Kordesch et al., 2008; Ping et al., 2009; Davis et al., 2009). Incorporation of long-distance drainage means that lacustrine sedimentation and paleoenvironments may have been influenced by climatic, tectonic, and geomorphic conditions at locations far removed from the lake itself. As an extreme example, the modern Caspian Sea receives drainage from up to 1000 km north in the Urals, whereas the adjacent Aral Sea receives drainage from over 1500 km southeast in the Pamir. The effect of anthropogenic drainage modification has been dramatically demonstrated for the Aral Sea, which has been reduced to little more than a playa by upstream reductions in river flow. Natural modification of ancient river courses due to tectonic, volcanic, or geomorphic changes could be expressed in similarly dramatic fashion, through facies changes in associated lacustrine strata.

The Green River Formation presents an ideal opportunity to test the importance of long-distance regional drainage in determining the character of ancient lacustrine deposits by using the extensive geologic framework that has already been erected for both it and the surrounding North American Cordillera and its foreland. In particular, recent geochronologic advances have made the Green River Formation one of the best-dated intervals of lacustrine strata in the world (Smith et al., 2003, 2008b, 2010). The

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present study focuses on the sudden introduction of volcaniclastic sediment at ca. 49.6 Ma, which is interpreted to reflect the expansion of the drainage basin feeding Lake Gosiute (Surdam and Stanley, 1980). It has long been assumed that the source of the volcanic sediment was the Absaroka volcanic province of northwestern Wyoming and southern Montana (Koenig, 1960; Ebens, 1963; Bradley, 1964; Surdam and Stanley, 1980), primarily because of the proximity of the Absaroka volcanic province to the greater Green River Basin and because of south-directed paleocurrents observed in volcaniclastic strata along the northern margin of the basin (Groll and Steidtmann, 1987). The Absaroka volcanic province, however, was not the only major volcanic system active in the region during the middle Eocene. The Challis volcanic field in central Idaho was active from 49.8 to 45.5 Ma (Fisher et al., 1992; Janecke and Snee, 1993; M’Gonigle and Dalrymple, 1996; Janecke et al., 2005) and produced volcanic rock and volcaniclastic sandstone that was similar in bulk chemical composition to rocks of the Absaroka volcanic province. Carroll et al. (2008) and Dobbert et al. (2010) proposed on the basis of ⁸⁷Sr/⁸⁶Sr and ⁴⁰Ar/³⁹Ar data in lacustrine carbonate facies that an abrupt change in sedimentation in the Washakie subbasin that occurred at ca. 49 Ma from balance-filled to overfilled lake conditions (cf. Carroll and Bohacs, 1999; Bohacs et al., 2000) occurred due to the sudden capture of a major volcaniclastic–bearing river system, upstream of the basin, which rose not in the Absaroka volcanic province, but rather in the Challis volcanic field, ~500 km away in central Idaho. The present study directly tests this hypothesis, based on detailed examination of regional geologic relationships in Idaho and on petrographic and radioisotopic analyses of the volcaniclastic detritus itself. Our results establish the source of volcaniclastic sandstone in the greater Green River Basin and permit discussion of the configuration of the Eocene landscape in the northwestern United States.

GEOLOGIC SETTING

The largest and longest-lived Eocene lakes of the western United States occupied the ponded basins discussed by Dickinson et al. (1988), which include the greater Green River Basin as well as the Piceance Creek, Uinta, Wind River, and Bighorn Basins (Fig. 1). These basins were impounded in the portion of the Sevier foreland that was segmented by Laramide basement-cored uplifts beginning in the Maastrichtian (Love, 1960; Dickinson et al., 1988; DeCelles, 2004). Deposition of the youngest lacustrine strata in the greater Green River Basin was coeval with late-stage Laramide deformation (Dickinson et al., 1988; Smith et al., 2008b; Chetel and Carroll, 2010), eruptions in upstream volcanic provinces (McIntyre et al., 1982; Janecke and Snee, 1993; Hiza, 1999; Feeley et al., 2002), and a fundamental, long-term shift from convergent to extensional tectonics (Janecke, 1994; Constenius, 1996; Foster et al., 2007).

The lacustrine strata of the Green River Formation in the greater Green River Basin is ensnared in alluvial clastic sedimentary rocks (Bradley, 1964; Roehlter, 1992). Alluvial deposits are differentiated into two classes on the basis of petrography: internally sourced quartzo-feldspathic sandstone, assigned to the Wasatch Formation (Denson and Pipirigos, 1969; Braungel and Stanley, 1977; Sullivan, 1985), and volcaniclastic sandstone (Surdam and Stanley, 1980; Dickinson et al., 1988) that was externally derived from a distant source, assigned to the Bridger, Washakie, and Uinta Formations (Matthew, 1909; Roehler, 1973b; Cashion and Donnell, 1974), as well as the Sand Butte Bed of the Laney Member of the Green River Formation (Roehler, 1973a).

Eocene Volcanism in the Western Interior United States

Thick-skinned, Laramide-type deformation in the western interior in the latest Cretaceous–Paleocene was notably amagmatic (cf. Dickinson and Snyder, 1978), although volcanic activity was voluminous in the Eocene between 55 and 40 Ma along a belt that stretched from Washington into Montana and Wyoming (Snyder et al., 1976; Armstrong and Ward, 1991). This belt includes the Clarno volcanic arc, the Absaroka volcanic province of Montana and Wyoming, and the Challis volcanic field of central Idaho.

Absaroka Volcanic Province

The volcanic and volcaniclastic rocks of the Absaroka volcanic province (Fig. 1) were erupted from more than 13 major volcanic centers that covered an area of ~23,000 km² in western Wyoming and southwestern Montana (Smedes and Prostka, 1972). A northwest to southeast progression of ages shows that the locus of volcanism shifted through time (Fig. 2). Absaroka volcanic rocks are divided into three groups based on age and composition (Smedes and Prostka, 1972). In ascending stratigraphic order, they are the Washburn, Sunlight, and Thorofare Creek Groups.

The rocks of the Washburn Group are restricted to the northern part of the field and are composed of hornblende and pyroxene andesite, and lesser amounts of biotite andesite and dacite, minor amounts of basaltic lavas, and rhyodacite ash-flow tuff (Smedes and Prostka, 1972). An age of 55.20 ± 0.61 Ma for a dacite flow at the base of the sequence (Feeley et al., 2002) suggests that volcanism may have commenced as early as 55 Ma (Fig. 2). Note: All ⁴⁰Ar/³⁹Ar ages cited in this study have been normalized to the intercalibration values of Renne et al. (1998), and are reported with 2σ analytical and intercalibration uncertainties. All ages are based on the ⁴⁰Ar/³⁹Ar clock unless otherwise noted.

Eruption of the overlying Sunlight Group is constrained between 49.83 ± 0.19 Ma and 48.37 ± 0.15 Ma (Feeley and Cosca, 2003). Sunlight Group rocks are distributed over a broader area than the Washburn Group and are composed of dark-colored pyroxene andesite lava flows, volcaniclastic rocks, and potassic basalts (Smedes and Prostka, 1972).

The youngest rocks of the Absaroka volcanic province belong to the Thorofare Creek Group and largely postdate deposition of volcaniclastic sandstone in the greater Green River Basin. The base of the Thorofare Creek Group overlaps the age of the youngest rocks of the Sunlight Group at ca. 48 Ma (Hiza, 1999). These rocks are composed of volcaniclastic strata and andesite lavas (Smedes and Prostka, 1972). The end of volcanism in the Absaroka volcanic province is constrained by an age of 43.95 ± 0.20 Ma (Hiza, 1999) for intrusive rocks at the southern end of the province.

Small volumes of intrusive rocks in the Crazy Mountains Basin, located north of the Absaroka volcanic province in west-central Montana (Fig. 1), are also Eocene in age. A ⁴⁰Ar/³⁹Ar age of 50.89 ± 0.18 Ma (Harlan, 2006) for a trachyte sill and a 49.47 ± 0.17 Ma diorite (Wilson and Elliott, 1997) constrain the ages of volcanic activity in the Crazy Mountains Basin. Volcanic rocks in the Crazy Mountains Basin include sodium-rich, alkaline mafic syenites and trachytes and subalkaline rocks consisting of potassium-rich diorites and granodiorites (du Bray and Harlan, 1996; Harlan, 2006).

Challis Volcanic Field

The Challis Volcanic Group of central Idaho (Fig. 1) preserves remnants between 42°N and 49°N, forming the largest region of Eocene volcanic rocks in the United States. Today, Challis volcanic rocks are preserved over an area of ~25,000 km²; however, if the Challis volcanic field is expanded to include areas where the Idaho Batholith is intruded by Eocene plutons, then the full spatial extent of volcanic activity expands to >100,000 km² (Link and Lewis, 2009). The calc-alkaline rocks of the Challis
Figure 1. Map showing the locations of Eocene nonmarine basins in the Sevier segmented foreland, compiled from Ross et al. (1955), Bond and Wood (1978), and Love and Christiansen (1985). Abbreviations: SRSZ—Salmon River shear zone, PB—Pioneer Batholith.
Figure 2. $^{40}$Ar/$^{39}$Ar geochronology for the Eocene volcanic rocks of Wyoming, Idaho, and Montana (gGRB—greater Green River Basin). Ages are relative to the standard values of Renne et al. (1998) and shown with 2σ intercalibration uncertainties. Also plotted are the Eocene $^{40}$Ar/$^{39}$Ar single-crystal ages for detrital K-feldspars from six samples from this study. The $^{40}$Ar/$^{39}$Ar ages are shown for Absaroka Volcanic Supergroup (Harlan et al., 1996; Hiza, 1999; Feeley et al., 2002; Feeley and Cosca, 2003); Challis Volcanic Group (Janecke and Snee, 1993; Snider, 1995; M’Gonigle and Dalrymple, 1996; Janecke et al., 1997; VanDenburg et al., 1998; Janecke et al., 1999; Janecke and Blankenau, 2003; Sanford, 2005); Crazy Mountains (Harlan, 2006); Dillon volcanics (Fritz et al., 2007, K-Ar); and Lowland Creek volcanics (Ispolatov, 1997). Note that some error bars are smaller than their symbol. The shaded bar indicates the interval of deposition of volcaniclastic sandstone in the basins of the segmented foreland.
Volcanic Group were erupted onto an irregular surface of Proterozoic quartzite, Paleozoic sedimentary rock, and the Cretaceous Idaho Batholith (Rodgers and Janecke, 1992). Stratigraphic relations within the Challis volcanic field are highly complex due to an irregular basal surface, as well as extensional faulting that occurred during and after Challis volcanic activity (McIntyre et al., 1982; Kiilsgaard et al., 1986, 2000; Snider, 1995; Janecke et al., 1997).

The Challis volcanic field is composed of (1) intermediate lava flows, fissure eruptions, and point source centers in the southeastern third of the field, (2) large calderas filled with 2–3-km-thick ash-flow tuffs, northeast-striking grabens, and trap door calderas in the central axis of the field, and (3) Eocene granites and granodiorites encased in the Cretaceous Idaho Batholith along the more deeply exhumed western margin (Hardyman, 1981, 1985; McIntyre et al., 1982; Leonard and Marvin, 1982; Ekren, 1985, 1988; Janecke and Snee, 1993; Snider, 1995; Janecke et al., 1997; Gaschnig et al., 2007; Link and Lewis, 2009; Skipp et al., 2009).

Activity in the Challis volcanic field is constrained by a suite of 40Ar/39Ar ages from several studies in the southeast part of the field and K-Ar and U-Pb data elsewhere (Fig. 2). These data reflect both effusive and explosive phases of activity in the Challis volcanic field. The early phase (ca. 50–47 Ma) was primarily effusive, with andesitic and dacitic lavas produced overlying rocks are primarily dacitic and andesitic rocks, with andesite and dacite lavas produced distributary of the underlying biotite-bearing tuffs (i.e., biotite ash-flow tuff in Muddy Creek Basin dated at 47.33 ± 0.30 Ma; Janecke et al., 1999) suggest that dacitic and andesite lava flows filled and leveled pre-Eocene erosional topography prior to eruption of the Tuff of Challis Creek (Janecke and Snee, 1993).

**Other Eocene Volcanism**

Volcanic activity immediately adjacent to the Challis volcanic field is documented in several other localities in the North American Cordillera during the middle Eocene. The Dillon volcanic field contains relatively thin and discontinuous units, and occupies the easternmost edge of the Challis volcanic field in western Montana. The lower Dillon volcanic rocks are temporally equivalent to the Challis volcanic rocks, are plotted as such in Figure 2, and are considerably more felsic than middle and upper Dillon volcanic rocks (Fritz et al., 2007). The Lowland Creek volcanics, farther to the north, are a sequence of quartz-latitic rocks that rest unconformably on the Cretaceous Boulder Batholith, northeast of the main Challis zone (Fig. 1; Smedes and Thomas, 1965). The 40Ar/39Ar ages (Ispolatov, 1997) indicate that the earliest volcanic activity in the Lowland Creek volcanic field predates the earliest activity in the Challis volcanic field by ~4 m.y. and overlaps temporally with volcanism in the Absaroka volcanic province (Fig. 2).

**VOLCANICLASTIC SEDIMENTATION IN THE EOCENE FORELAND**

Sandstone that has been described in the past as volcanioclastic is observed in several Laramide basins beginning in the middle Eocene (e.g., Van Houten, 1964; Love, 1970; Surdam and Stanley, 1980; Johnson, 1981). The stratigraphic and temporal constraints on these units, as well as petrographic analyses, are summarized next. Although stratigraphic and temporal constraints are highly variable, a basic understanding of these constraints is critical to the interpretation of paleodrainage patterns in the middle Eocene.

**Greater Green River Basin**

Volcaniclastic sandstone was deposited in the greater Green River Basin between 49.6 Ma and 47 Ma (Fig. 3). This interval is constrained by the age of the Sixth Tuff (49.58 ± 0.17 Ma; Smith et al., 2008b) and the age for the youngest dated material in the greater Green River Basin, the Sage Creek Mountain pumice (47.14 ± 0.16 Ma; Smith et al., 2008b), which has a biostratigraphic age of Bridge E (ca. 47 Ma; Fig. 3). Volcaniclastic sandstones are assigned to three units, the alluvial/fluvial Bridge and Washakie Formations, as well as the deltaic Sand Butte Bed of the Laney Member of the Green River Formation (Culbertson, 1962; Riehler, 1973a, 1973b; Chelet and Carroll, 2010). In the greater Green River Basin, volcanioclastic sandstone is temporally equivalent to the lacustrine Laney Member; the oldest volcanioclastic sandstone occurs between the transition between underfilled (Wilkins Peak Member) and balance-filled (lower Laney Member) strata and the overlying transition to overfilled strata (upper Laney Member) in the Bridge subbasin (Smith et al., 2008b; Chelet and Carroll, 2010). High-resolution correlation indicates that the oldest volcanioclastic sandstone is closer in age to the underfilled to balance-filled transition and that volcanioclastic sandstone progressively infilled the greater Green River Basin from northwest to southeast. Point counts of 41 thin sections from the Bridge and Washakie Formation as well as the Sand Butte Bed show that volcanioclastic sandstone in the greater Green River Basin has a composition that is consistent with a magmatic source (Fig. 4; see GSA Data Repository Table DR1 for raw data¹). These samples belong to the volcanioclastic petrofacies described in Chelet and Carroll (2010). The most common lithic fragment is volcanic, the plutonic/volcanic ratio is low, there is abundant biotite and hornblende, and muscovite is absent.

**Piceance Creek Basin**

Volcanioclastic sandstone was deposited in the Piceance Creek Basin beginning ca. 48.4 Ma (Fig. 3). This is constrained by the age for the Wavy Tuff, which is ~30 m above the Mahogany zone in the Piceance Creek and Uinta Basins (48.33 ± 0.27 Ma; Remy, 1992; Smith et al., 2008b). The age of the youngest volcanioclastic-bearing strata is more difficult to constrain, but it is estimated to have a biostratigraphic age of late Bridger-3 or early Uintan-1 (Fig. 3; Smith et

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¹GSA Data Repository item 2010213, Raw point count data, full documentation of 40Ar/39Ar geochronology, and documentation of existing 40Ar/39Ar and U-Pb geochronology for potential source rocks and supporting references, is available at http://www.geosociety.org/pubs/ft2010.htm or by request to editing@geosociety.org.
Figure 3. Composite stratigraphic sections, age model, and biostratigraphic zonation for basins that preserve Eocene sediment that has been described in the past as volcaniclastic. The age model is based on the work of Smith et al. (2008b). The sampled interval is highlighted; more detailed information regarding the stratigraphic positions of each sample can be found in Figure DR1 (see text footnote 1). Abbreviations: gGRB—greater Green River Basin, WRB—Wind River Basin, PCB—Piceance Creek Basin.

Figure 4. Volcaniclastic sandstone compositional data for sandstone samples from three basins plotted on a QFL and ternary diagram. Composition means are plotted, and 1σ standard deviations are indicated by the patterned polygons. Shaded provenance fields are from Dickinson and Suczak (1979). Symbols: squares—greater Green River Basin, triangles—Piceance Creek Basin, circles—Wind River Basin. Abbreviations: Qt—total quartz, F—feldspar, L—lithic fragments, gGRB—greater Green River Basin, WRB—Wind River Basin, PCB—Piceance Creek Basin.
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The most common lithic fragments in these samples are polycrystalline and graphitic quartz, microcline is abundant, and the plutonic:volcanic ratio is higher than that of the samples from the greater Green River Basin and Piceance Creek Basin (Fig. 4). It therefore appears unlikely that the Wagon Bed Formation is genetically related to volcaniclastic deposits associated with the Green River Formation.

ISOTOPIC RESULTS

Pb Isotopes

There are significant differences in the Pb isotope composition of volcanic rocks from the Absaroka and Challis volcanic fields (Peterman et al., 1970; Meen and Eggler, 1987; Norman and Leeman, 1989; Norman and Mertzman, 1991; Hiza, 1999). Absaroka volcanic rocks tend to have lower 206Pb/204Pb and 208Pb/204Pb values as compared to rocks from the Challis volcanic field (Fig. 5). There is some overlap in 206Pb/204Pb between Challis and Absaroka volcanic rocks, but for most rocks that have these overlapping 206Pb/204Pb ratios, the Absaroka volcanic rocks tend to have higher 207Pb/204Pb values (Fig. 5); this likely reflects the old
greater than the Challis volcanic field rocks, it is expected that some of the detritus was derived from sources that match the composition of Idaho Batholith rocks (Fig. 5), located to the north and west of the Challis volcanic field.

**40Ar/39Ar Geochronology**

We obtained a total of 256 K-feldspar apparent ages from six samples of volcaniclastic sandstone from the greater Green River Basin (Table 1). The most prominent population is Cenozoic in age; however, Mesozoic (23% of the grains) and Proterozoic (32% of the grains) populations are also present. These results are interpreted through comparison to a compilation of published 40Ar/39Ar, K-Ar, and U-Pb data from the rocks of the Idaho–Montana segment of the North American Cordillera (see Tables DR3 and DR4 [footnote 1]). Comparison of Eocene and Mesozoic ages provides a high degree of confidence when inferring sources, whereas interpretation of Proterozoic ages is less certain because the majority of Neoproterozoic and Mesoproterozoic tectonic and magmatic events are based on U-Pb geochronology. Thus, the feldspar cooling ages of these old sources are not well known.

These results show that juvenile volcanic sediment was mixed with grains derived from other prevolcanic sources. The youngest detrital ages are broadly consistent with the distribution of previously reported ages of Challis volcanic rocks (Fig. 2). Conversely, ages similar to base- men exposed within the foreland (1.6–1.8 Ga; Premo and VanSchmus, 1989) are absent.

**THE EOCENE LANDSCAPE**

Four major rock types were exposed in central Idaho and western Montana during the Eocene (Fig. 8). Eocene volcanic rocks overlie Paleozoic sedimentary rock as well as the granitic rocks of the Cretaceous Idaho Batholith and Proterozoic quartzite. The prominent Eocene, Cretaceous, and Proterozoic populations that are observed in the detrital K-feldspar data (Fig. 6) are consistent with the interpreted Eocene geology. We note that Paleozoic sedimentary rocks are primarily limestone, which would not have contributed detrital feldspars to greater Green River Basin sandstone.

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**TABLE 1. Pb ISOTOPE COMPOSITIONS OF GREATER GREEN RIVER BASIN SANDSTONE**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Latitude (°N)</th>
<th>Longitude (°W)</th>
<th>206Pb/204Pb</th>
<th>207Pb/204Pb</th>
<th>208Pb/204Pb</th>
</tr>
</thead>
<tbody>
<tr>
<td>05-CP-154</td>
<td>42°16.2'</td>
<td>108°43.7'</td>
<td>18.371 ± 0.002</td>
<td>15.939 ± 0.002</td>
<td>38.754 ± 0.004</td>
</tr>
<tr>
<td>05-CP-107.4</td>
<td>42°16.2'</td>
<td>108°43.7'</td>
<td>18.119 ± 0.005</td>
<td>15.586 ± 0.005</td>
<td>38.649 ± 0.018</td>
</tr>
<tr>
<td>05-CP-107.4*</td>
<td>42°16.3'</td>
<td>108°43.7'</td>
<td>18.099 ± 0.003</td>
<td>15.560 ± 0.002</td>
<td>38.562 ± 0.006</td>
</tr>
<tr>
<td>05-CP-21</td>
<td>42°16.2'</td>
<td>108°43.7'</td>
<td>18.528 ± 0.006</td>
<td>15.603 ± 0.006</td>
<td>38.839 ± 0.018</td>
</tr>
<tr>
<td>05-CP-1</td>
<td>42°16.2'</td>
<td>108°43.7'</td>
<td>18.532 ± 0.002</td>
<td>15.608 ± 0.002</td>
<td>38.952 ± 0.004</td>
</tr>
<tr>
<td>05-CP-105.1</td>
<td>41°16.7'</td>
<td>108°43.6'</td>
<td>18.668 ± 0.002</td>
<td>15.697 ± 0.002</td>
<td>38.816 ± 0.005</td>
</tr>
<tr>
<td>05-CP-105.5</td>
<td>41°16.7'</td>
<td>108°43.8'</td>
<td>18.680 ± 0.002</td>
<td>15.688 ± 0.002</td>
<td>38.774 ± 0.004</td>
</tr>
<tr>
<td>07-FT-2</td>
<td>107°51.5'</td>
<td>19.629 ± 0.002</td>
<td>15.717 ± 0.001</td>
<td>39.292 ± 0.003</td>
<td></td>
</tr>
<tr>
<td>07-FT-3</td>
<td>107°51.6'</td>
<td>19.627 ± 0.002</td>
<td>15.716 ± 0.002</td>
<td>39.284 ± 0.005</td>
<td></td>
</tr>
<tr>
<td>05-CPI</td>
<td>42°16.2'</td>
<td>108°43.8'</td>
<td>17.372 ± 0.002</td>
<td>15.538 ± 0.002</td>
<td>37.373 ± 0.005</td>
</tr>
<tr>
<td>05-KR-1</td>
<td>41°6.8'</td>
<td>108°51.0'</td>
<td>18.014 ± 0.002</td>
<td>15.559 ± 0.002</td>
<td>38.523 ± 0.004</td>
</tr>
<tr>
<td>05-SB-108.2</td>
<td>42°20.8'</td>
<td>108°40.10'</td>
<td>17.971 ± 0.002</td>
<td>15.536 ± 0.002</td>
<td>38.557 ± 0.005</td>
</tr>
<tr>
<td>05-GD-130.9</td>
<td>41°6.3'</td>
<td>108°32.2'</td>
<td>18.242 ± 0.001</td>
<td>15.568 ± 0.001</td>
<td>38.695 ± 0.003</td>
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<tr>
<td>05-GD-130.9*</td>
<td>41°6.3'</td>
<td>108°32.2'</td>
<td>18.241 ± 0.002</td>
<td>15.568 ± 0.002</td>
<td>38.745 ± 0.006</td>
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<tr>
<td>05-AC-3</td>
<td>41°23.5'</td>
<td>108°30.5'</td>
<td>18.029 ± 0.002</td>
<td>15.562 ± 0.005</td>
<td>38.514 ± 0.013</td>
</tr>
<tr>
<td>01-AC-3*</td>
<td>41°23.5'</td>
<td>108°30.5'</td>
<td>17.998 ± 0.002</td>
<td>15.524 ± 0.001</td>
<td>38.383 ± 0.004</td>
</tr>
<tr>
<td>06-WM-1</td>
<td>41°32.6'</td>
<td>109°7.1'</td>
<td>18.051 ± 0.003</td>
<td>15.559 ± 0.003</td>
<td>38.688 ± 0.008</td>
</tr>
<tr>
<td>06-WM-1*</td>
<td>41°32.6'</td>
<td>109°7.1'</td>
<td>18.046 ± 0.003</td>
<td>15.553 ± 0.003</td>
<td>38.662 ± 0.007</td>
</tr>
<tr>
<td>06-DR-6</td>
<td>42°10.0'</td>
<td>110°8.4'</td>
<td>18.547 ± 0.002</td>
<td>15.501 ± 0.002</td>
<td>38.377 ± 0.004</td>
</tr>
<tr>
<td>07-WP-1</td>
<td>41°29.4'</td>
<td>109°20.4'</td>
<td>18.184 ± 0.002</td>
<td>15.555 ± 0.002</td>
<td>38.590 ± 0.005</td>
</tr>
<tr>
<td>06-DR-6</td>
<td>42°10.0'</td>
<td>110°8.4'</td>
<td>20.171 ± 0.004</td>
<td>15.768 ± 0.002</td>
<td>39.364 ± 0.009</td>
</tr>
<tr>
<td>05-GR-2</td>
<td>41°29.4'</td>
<td>109°27.5'</td>
<td>18.030 ± 0.002</td>
<td>15.564 ± 0.002</td>
<td>38.840 ± 0.004</td>
</tr>
<tr>
<td>05-GR-2*</td>
<td>41°29.4'</td>
<td>109°27.5'</td>
<td>18.020 ± 0.002</td>
<td>15.552 ± 0.002</td>
<td>38.791 ± 0.004</td>
</tr>
</tbody>
</table>

**Note:** The errors are 2σ based on in-run statistics for each Pb isotope analysis (50 s integrations with a 208Pb ion current of 2 × 10–11 amps). Bulk-rock samples from the greater Green River Basin (gGRB) were dissolved in HF and HNO3, and Pb was separated using HBr–HCl anion exchange columns. Total procedural blanks ranged from 99 to 128 pg of Pb. Pb isotope ratios were measured by thermal ionization mass spectrometry using Re filaments and Si-gel on a Micromass Sector 54 instrument at the University of Wisconsin–Madison Radiogenic Isotope Laboratory. Repeat analyses of NIST SRM-981 (9 analyses) and SRM-982 (8 analyses) were used to correct for instrumental mass bias, where the pooled average mass dependent fractionation factor determined from the measured 207Pb/204Pb and 208Pb/204Pb ratios of NIST SRM-981 and SRM-982, respectively, was 0.0014 ± 0.0008 (2 S.D.) per amu. Eight bulk rocks were analyzed twice, and reproducibility was within the uncertainty of the empirically determined fractionation factor. The stratigraphic positions of all analyzed samples are illustrated in Figure DR1 (see text footnote 1).

**Duplicate Pb isotope analyses on the same dissolution.**
Prior to the eruption of the Challis volcanic field, Middle Proterozoic rocks at the surface in the north transitioned to progressively younger strata to the west, south, and east (Rodgers and Janecke, 1992). Challis volcanic rocks were erupted onto a dissected surface, which featured a network of preexisting paleovalleys (Janecke et al., 2000). In the southeast part of the Challis volcanic field, many complete sections of Challis volcanic rocks are preserved in numerous late Eocene to early Miocene half grabens. Calderas in the central axis of the field also preserve complete sections, but exposure of the older units is limited, except in tilted trap-door calderas (Ekren, 1988; Janecke et al., 1997). In contrast, the western and northern third of the volcanic field has been denuded of its volcanic cover, and the plutonic roots are exposed in the eastern part of the Idaho Batholith.

The western third of the volcanic field is the most likely source of voluminous volcaniclastic sediment. The documented paleovalleys, which are all in the southeast part of the Challis volcanic field, preserve thicker sections of middle Eocene volcanic and sedimentary rocks that include shoestring deposits of unwelded ash-fall tuff. The 49.34 ± 0.28 Ma quartzite-bearing tuff (M’Gonigle and Dalrymple, 1996), which forms a shoestring deposit that is 85 km long and up to 5 km wide at the base of the Lemhi Pass paleovalley (Janecke et al., 2000), the lithic-rich tuff, also preserved in the Lemhi Pass paleovalley, and the 47.87 ± 0.20 Ma Tuff of Curtis Ranch, are poorly welded and today are mostly confined to the paleovalley and the source calderas (Blankenau, 1999; Janecke and Blankenau, 2003). Based on these observations, we infer that, at times, the Eocene landscape was blanketed with easily eroded pyroclastic rocks, most of which have been subsequently eroded.

The Idaho Batholith is composed of two temporally and geographically distinct lobes. The older Atlanta Lobe intruded Paleozoic sedimentary rock west of the Challis volcanic field and is both overlain and intruded by Eocene rocks (Link and Lewis, 2009). The northern Bitterroot Lobe invaded Proterozoic metasedimentary rocks, is heavily intruded by Eocene plutons (Toth and Stacey, 1992), and hosted three Eocene metamorphic core complexes that formed as the region underwent extensional collapse (Foster et al., 2007). Thermochronology from the Bitterroot metamorphic core complex indicates that extension was initiated between 54 and 52 Ma (Foster et al., 2001), although rapid exhumation did not begin until ca. 50 Ma (Foster and Fanning, 1997; Foster et al., 2001; House et al., 2002), and continued until after 40 Ma (Fig. 9; Foster et al., 2007). Total Eocene exhumation was likely ~20–25 km in the eastern part of the

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**Figure 6.** Comparison of the ⁴⁰Ar/³⁹Ar probability density plots of the six sandstone samples from the greater Green River Basin analyzed in this study. The profiles are in approximate stratigraphic order, with the oldest at the bottom. Data for sample 01AC3 are from Smith et al. (2008b). Each sample is plotted at two scales. The histograms illustrate the full distribution of ages, while the probability density curves highlight the Cenozoic and Mesozoic age distributions.
core complex (Foster et al., 2001) and ~10 km in the western part of the core complex (House et al., 2002), which generated a large volume of detritus that was not locally preserved. Eocene Cordilleran metamorphic core complexes were associated with a high-elevation landscape; the elevation in the Shuswap metamorphic core complex in southern British Columbia is estimated to have been more than 4 km (Mulch et al., 2004, 2007) based on hydrogen isotope data. In contrast, a floral-assemblage–based estimate of elevation in the Thunder Mountain caldera in central Idaho is estimated to have been only 1.7 km (Axelrod, 1998).

SEDIMENT PROVENANCE INTERPRETATIONS

Eocene ages from K-feldspar in the greater Green River Basin range from 47.5 Ma to 52.3 Ma and overlap activity in both the Challis and Absaroka volcanic fields and the intervening Lowland Creek volcanic center (Figs. 2 and 7). The bulk of the Eocene ages (n = 106, ~92% of Eocene ages) overlap with the abbreviated interval of Challis volcanism, which began ca. 50 Ma, 5.5 m.y. after initial volcanism in the Absaroka volcanic province (Fig. 2). The primary source of Eocene K-feldspar grains preserved in the greater Green River Basin is interpreted as the Challis volcanic field on the basis of the Pb isotope results discussed here and the Eocene cooling ages (Figs. 2 and 5). Now-eroded volcanic rocks that once blanketed the landscape in the northern and western parts of the volcanic field are the likely source of this material and account for abundant biotite and hornblende in greater Green River Basin sandstone samples.

Eocene ages that slightly predate initial volcanism in the Challis volcanic field (n = 9, ~8% of the Eocene ages; Figs. 2 and 9) may have had a volcanic source in the Absaroka volcanic province, the Lowland Creek volcanics, undated parts of the Challis volcanic field, or Eocene metamorphic core complexes at the north and east edge of the Challis volcanic field. Simple mixing calculations between Absaroka and Challis Pb reservoirs demonstrate that for the majority of samples, if Absaroka-derived Pb was present, it was likely less than 25% of the total (Fig. 5). Because the footwalls of Eocene metamorphic core complexes in Idaho and western Montana experienced as much as 25 km of unroofing between 53 and 40 Ma (Fig. 9; House et al., 2002; Foster et al., 2007), they constitute another possible source for sediment that would yield pre-Challis and Challis Eocene ages.

Cretaceous and earliest Paleocene ages form the second most prominent population of the cooling ages (n = 36, ~14% of total; Fig. 7). These ages are consistent with cooling ages from the northern part of the Idaho Batholith (Table DR1 [see footnote 1]). The 65 Ma mean age of this population overlaps the mean cooling age for the Bitterroot Lobe in north-central Idaho (Criss and Fanning, 1993; Foster and Fanning, 1997). The Atlanta Lobe, in central Idaho, has a mean cooling age of 76 Ma (Lewis et al., 1987), and a population of grains of this mean age is notably absent from the six sandstone samples analyzed in this study (Fig. 7). The rare occurrence of muscovite in greater Green River Basin sandstone (Table DR1 [see footnote 1]) precludes large volumes of sediment derived from the two-mica granite and granodiorite of the Bitterroot Lobe (Toth and Stacey, 1992; Foster and Fanning, 1997), which is in agreement with the limited number of Cretaceous cooling ages; hydrologic winnowing may have removed some muscovite from the depositional system. White mica, however, is common in sedimentary rocks of the same age, such as the Tyee Formation, which were deposited on the western side of the North American Cordillera. White mica grains with ages consistent with the mean cooling age for the Bitterroot Lobe are present in the Tyee Formation (Heller et al., 1992).

A comparatively small percentage of the detrital K-feldspar grains (n = 81, ~32% of total ages) analyzed as part of this study yielded Proterozoic ages between 622 Ma and 1.5 Ga. There are no Archean cooling ages and only one Paleoproterozoic 2.4 Ga cooling age. It should be noted, however, that effort was made to select for unaltered, volcanicogenic Eocene grains, and, as a result, Proterozoic populations are likely underrepresented. Potentially widespread sources of Neoproterozoic and Mesoproterozoic cooling ages include metamorphosed Belt Supergroup (ca. 1.1 Ga; Obradovich and Peterman, 1968; Doughty and Chamberlain, 2007), Neoproterozoic 685 ± 7 Ma sedimentary rocks, intrusions in the west Salmon River Arch area (Lund et al., 2003), and post-Belt Supergroup
anorogenic granite and gabbro (ca. 1.4 Ga; Doughty and Chamberlain, 1996, 2007; Table DR2 [see footnote 1]). Syn–Belt Supergroup anorogenic granite and gabbro (ca. 1.4 Ga; Dragovich et al., 2009) could have been recycled from pre-Eocene volcanic rocks (ca. 1.5–1.4 Ga) are so small in volume that they are unlikely sources of sediment to the greater Green River, and Piceance Creek Basins with volcanioclastic sediment derived from the Absaroka volcanic province (Surdam and Stanley, 1980). Recognition of a significant volume of Challis-derived sediment in the detrital feldspars (Kellogg et al., 2003). The Eocene Hawley Creek paleovalley flowed through the center of the impact area (Fig. 8; Ja- necke et al., 2000).

**IMPLICATIONS FOR PALEODRAINAGE**

The prevailing model for delivery of volcanioclastic sediment to the greater Green River Basin was the progressive infill of the Wind River, greater Green River, and Piceance Creek Basins with volcanioclastic sediment derived from the Absaroka volcanic province (Surdam and Stanley, 1980). Recognition of a significant volume of Challis-derived sediment in the foreland necessitates revision of this model. We propose a connection between an orogenic river

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**Note:** The data from sample 01AC3 are not included here because they are already published in Smith et al. (2008b). Feldspar crystals between 200 and 500 microns in length were isolated using the methods described by Smith et al. (2003) and irradiated together with flux monitors at the Oregon State University Triga reactor. Ar isotopic compositions were determined from single crystals via automated CO2 laser degassing and mass spectrometry in single crystal total fusion experiments (according to the procedure of Smith et al., 2000). The 28.34 Ma Taylor Creek sanidine was used as the neutron fluence standard (Renne et al., 1998) against which ages for individual experiments were calculated using the decay constant of Steiger and Jäger (1977).
with headwaters in the Idaho-Montana segment of the North American Cordillera that flowed through the southwest Montana re-entrant in the fold-and-thrust belt and entered the greater Green River Basin through its northwest corner. We name this drainage system the Idaho River (Fig. 8). The proposed Idaho River flowed ESE along the two paleovalleys of Janecke et al. (2000), with the larger and more northern Lemhi Pass paleovalley being the main trunk stream. Both paleovalleys are filled with Eocene volcanic rocks and compositionally diverse conglomerate derived from distant western source areas (Fig. 8). Cut-and-fill relations between units indicate that a river occupied the paleovalleys between the eruption of the oldest ash-flow tuff (dated at 49.78 ± 0.31 Ma; Janecke et al., 1999) and the overlying ca. 46 Ma Tuff of Challis Creek.

The distribution of Cretaceous detrital ages constrains the position of the continental divide during the Eocene east of the Atlanta Lobe of the Idaho Batholith and within or west of the Bitterroot Lobe (Fig. 8). As shown already, erosional products of the Bitterroot Lobe were transported east toward the greater Green River Basin, and ultimately south into the Piceance Creek Basin. The absence of either a 76 Ma population of detrital K-feldspar (Fig. 7) or granitic clasts in pre- to early Challis conglomerates deposited immediately east of the Atlanta Lobe (Janecke et al., 2000) suggests that the erosional products of the Atlanta Lobe were transported westward during the middle Eocene. The middle Eocene Tyee Formation in the Oregon Coast Ranges is the probable depocenter for this sediment (Heller et al., 1992). The up-section changes in the composition of sedimentary rocks from volcaniclastic to arkosic and a prominent population of 75 Ma white mica in detrital spectra from the Tyee Formation suggest that the Challis Lobe of the Idaho Batholith and its now-eroded carapace of Challis volcanic rocks were the dominant source of sediment (Chan and Dott, 1983; Heller et al., 1992). The youngest detrital ages from the Tyee Sandstone (Heller et al., 1992) establish deposition coeval with deposition of volcaniclastic sandstone in the greater Green River Basin.

We infer that the Idaho River drained toward the southeast until it reached the southwest Montana re-entrant. This divide is a full degree of longitude east of the drainage divide that separated east- and west-draining paleovalleys in the Nevada segment of the North American Cordillera beginning ca. 46 Ma and continuing into the Miocene (Henry, 2008). The major bend needed to align the continental divide of Henry (2008) with the Eocene one in central Idaho (Janecke et al., 2000) coincides with the major original bend in the Proterozoic rifted margin and the Cretaceous batholith belt (Lund, 2008).

The path between southwest Montana and the northern end of the greater Green River Basin is poorly constrained due to the widespread cover of the Eastern Snake River Plain. The path may have passed close to the Absaroka volcanic province (Fig. 8); however, the detrital K-feldspar age data and bulk-rock Pb isotope compositions provide clear evidence of Challis-derived juvenile volcanic detritus in greater Green River Basin volcaniclastic sandstone samples and little evidence for a significant amount of Absaroka-derived detritus. There are two possible explanations for the absence of Absaroka-derived Pb in greater Green River Basin volcaniclastic sandstone. The first is that a major drainage divide separated the Absaroka volcanic province from the Lake Gosuite watershed, and the other is that relatively little Absaroka volcanic rock has been eroded in comparison to the Challis volcanic field. We hypothesize that the Washakie Range acted as a drainage divide that separated the Absaroka volcanic province from the Lake Gosuite watershed (Fig. 8). The Washakie Range, in conjunction with the Wind River Mountains and the Granite Mountains, would have isolated the Absaroka volcanic province and the Wind River and Bighorn Basins from the Lake Gosuite watershed, which is consistent with the petrographic composition of temporally equivalent sandstone strata from each basin (Fig. 4). Although abundant south-oriented paleocurrents are observed in Eocene strata in the greater Green River Basin at the southeastern end of the Wind River Mountains (Groll and Steidtmann, 1987), they are limited to strata with a depositional age of ca. 47 Ma, which is younger than any of the samples analyzed as part of this study. Prior to the uplift of the Washakie Range, drainage from the Cordillera traveled east toward the Bighorn Basin (Sears and Ryan, 2003). An up-section shift in paleocurrent direction from east to south within the Paleocene Pinyon Conglomerate (Lindsey, 1972) is interpreted to reflect uplift of the Washakie Range and diversion of Cordilleran drainage south into the Lake Gosuite watershed. Eocene quartzite clast conglomerates in the Bighorn Basin are interpreted to have been recycled from Pinyon Conglomerate deposits that formed the crests of the Washakie and Basin Creek uplifts (Kraus, 1985) and document the continued high topographic relief of the range.

The uplift of the Washakie Range not only deflected Cordilleran-derived drainage south into the Lake Gosuite watershed, but it was also one of several Laramide uplifts that impounded a subsiding synformal basin beneath what is now the Absaroka volcanic province (Fig. 8; Sundell, 1990). Although the presence of the Absaroka Basin is masked today by the high-relief volcanic landscape of the Absaroka volcanic province, surficial geology and seismic and drill-core data are consistent with the presence of a relatively shallow northwest-trending basin (Sundell, 1990).

Prior to eruption of the Absaroka volcanic province, the Absaroka Basin was likely a closed basin. Lacustrine and paludal rocks assigned to the Willwood Formation are observed in association with basal volcanic units at several locations (Torres, 1985). The oldest sedimentary unit in the Absaroka Basin that is not deformed over the tops of Laramide anticlines is the Upper Wiggins Formation. The depositional age of the Wiggins Formation postdates the deposition of volcaniclastic sandstone in the greater Green River Basin (Sundell, 1990; Smith et al., 2008b), which indicates that the basin continued to subside until after the deposition of the
Figure 9. Detrital geochronology from this study includes Eocene ages that predate volcanism in the Challis volcanic field. There are several alternative sources for pre-Challis detrital ages, including metamorphic core complexes, the Absaroka volcanic province, and smaller Eocene volcanic zones. The histograms display the distribution of existing $^{40}$Ar/$^{39}$Ar and K-Ar geochronology for metamorphic core complexes and Eocene volcanic systems. The bin width for the histograms is 200 k.y. There are 132 $^{40}$Ar/$^{39}$Ar and K-Ar ages shown for the Eocene Anaconda (Montana; O’Neill et al., 2004; Foster et al., 2007), Bitterroot (Idaho-Montana; House and Hodges, 1994; Foster and Fanning, 1997; Foster et al., 2001, 2007; House et al., 2002), Clearwater (Idaho; Foster et al., 2007), Priest River (Washington-Idaho; Miller and Engels, 1975; Doughty and Price, 2000), and Shuswap (British Columbia; Mathews, 1981; Vanderhaeghe et al., 2003) metamorphic core complexes. The source of the data for Eocene volcanic rocks can be found in the caption for Figure 2. The superimposed probability density curve represents the distribution of Eocene ages from this study. All $^{40}$Ar/$^{39}$Ar ages have been intercalibrated to the values of Renne et al. (1998). The shaded bar indicates the interval of deposition of volcanioclastic sandstone in the greater Green River Basin. Ages for source rocks that lie to the left of the shaded bar cannot have contributed to the sandstones, but they are shown for completeness.
youngest volcaniclastic strata in the greater Green River Basin.

Eruption of Absaroka volcanic rocks into a subsiding northwest-trending synformal ba-
sin would have inhibited erosion and favored preservation; volcaniclastic detritus is common in the Absaroka volcanic province and formed large debris aprons that were preserved around volcanic centers (Smedes and Prostka, 1972). In contrast, little middle Eocene sediment is pre-
served in the Challis volcanic field outside of the Eocene paleovalleys. The volcanic rocks of the northern part of the Challis volcanic field were erupted onto a landscape that experienced con-
tinued development of topographic relief due to the development of Eocene metamorphic core complexes (Foster et al., 2007; Mulch et al., 2007) in the northern Cordillera. Erosional de-
nudation exhumed the roots of the volcanic field in the west. The ongoing uplift in the northern and western half of the Challis volcanic field is consistent with the deeper level of exposure there and limited preservation of volcanic rocks (Link and Lewis, 2009).

The absence of a significant volume of Ab-
saroka-derived sediment in greater Green River Basin volcaniclastic sandstone is explained either by a drainage divide or a structural and topographic rim around the Absaroka volca-
nic province. The most likely explanation is a combination of the two. While the Washakie Range restricted dispersal of Absaroka-derived sediment to the west and south, the eruption of Absaroka volcanic rocks into a closed drainage basin favored local preservation of Absaroka-
derived sediment.

CONCLUSIONS

This study documents a link between a source of water and sediment in central Idaho and non-
marine strata in the foreland. Lead isotope mea-
surements of volcaniclastic sandstone samples from the greater Green River Basin indicate that volcaniclastic sandstone in the greater Green River Basin originated from the Challis volcanic field and not the Absaroka volcanic province as previously thought. This result is consistent with the overlap between cooling ages of K-feldspar grains isolated from these samples and the short ca. 50–45 Ma interval of Challis magmatism, and not the longer 55–44 Ma Absaroka volcanic province. Pre-Eocene populations demonstrate that the Challis volcanic field was not the only source of greater Green River Basin sediment. The smaller northern Bitterroot Lobe also provided sediment to the basin, based on a mean age of 65 Ma for Cretaceous–earliest Paleocene detrital K-feldspar grains. The much larger At-
lanta Lobe flooded the area that became the Ore-
gon Coast Range at the same time, and was also being denuded of its Eocene volcanic cover and Cretaceous batholithic basement. We infer that the volcaniclastic sediment now in the greater Green River Basin originated primarily from the northern and western part of the Challis volcanic field. These areas have been denuded of their original volcanic cover, preserve the deep plutonic roots of the volcanic field where they intruded the Idaho Batholith, lie up gradi-
ent of the Eocene paleovalleys inferred to form the trunk stream of the Idaho River, and were demonstrably forming significant topographic relief during middle Eocene core-complex-
related extension. The central and southeast parts of the Challis volcanic field are still largely intact, except for tilting, burial, and erosion during younger extensional events. Further work is needed to confirm our preliminary interpre-
tation that subsidiary components in greater Green River Basin volcaniclastic sandstone were delivered to the Idaho River by tributaries that crossed the large 1.4 Ga anorogenic plutons aligned along the axis of the Salmon River Arch, crossed the center of the Neoproterozoic Be-
averhead impact site of southwest Montana, and tapped areas of Grenville-aged metamorphism north of the Bitterroot Lobe of the Idaho Batho-
lith. Zircon geochronology is in progress to fur-
ther test these hypotheses.

This result necessitates revision of the model for the river that delivered volcaniclastic sedi-
ment to the greater Green River Basin. Although the inferred path of the river passed close to the Absaroka volcanic province, significant mixing between Challis- and Absaroka-derived sediment was minimal because little Absaroka-
derived sediment was deposited outside the Absaroka Basin. Furthermore, the Idaho River connected the footwall of large Eocene meta-
orphic core complexes with a distant foreland, likely due to tsumescence during magmatism, inflow of middle or lower crust, and creation of new topographic highlands in the major exten-
sional terrains.

The results of this study help to demonstrate that regional drainage relationships exerted a primary influence on paleoenvironments and sedimentation in the Green River Formation and likely on other major lake systems. The abrupt shift toward permanently freshwater conditions in Lake Gosiute might easily be misinterpreted as indicating a climate change, whereas this shift can also be explained by the capture of a river draining higher elevations, upstream of the ba-
sin. Carroll et al. (2008) and Davis et al. (2009) suggested that the simultaneous deepening of Lake Uinta and deposition of the Mahogany oil shale across the Piceance Creek and Uinta Ba-
sins (in Colorado and Utah) occurred because of water spilling over from Lake Gosiute. If so, then deposition of one of the world’s richest and best known oil shale resources may be linked not to climate change, but instead to increased magmatism, crustal flow, and extension nearly 1000 km away.

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