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ABSTRACT

Shallow subduction of the Farallon plate beneath the western United States has been commonly accepted as the tectonic cause for the Laramide deformation during Late Cretaceous through Eocene time. However, it remains unclear how shallow subduction would produce the individual Laramide structures. Critical information about the timing of individual Laramide uplifts, their paleoelevations and paleoclimate at the time of uplift, and the temporal relationships among Laramide uplifts have yet to be documented at regional scale to address the question and evaluate competing tectonic models. The Wind River Basin in central Wyoming is filled with sedimentary strata that record changes of paleogeography and paleoelevation during Laramide deformation. We conducted a multidisciplinary study of the sedimentology, detrital zircon geochronology, and stable isotopic geochemistry of the lower Eocene Indian Meadows and Wind River formations in the northwestern corner of the Wind River Basin in order to improve understanding of the timing and process of basin evolution, source terrane unroofing, and changes in paleoelevation and paleoclimate.

Depositional environments changed from alluvial fans during deposition of the Indian Meadows Formation to low-sinuosity braided river systems during deposition of the Wind River Formation. Paleocurrent directions changed from southwestward to mainly eastward through time. Conglomerate and sandstone compositions suggest that the Washakie and/or western Owl Creek ranges to the north of the basin experienced rapid unroofing. The main body of the Washakie and Owl Creek ranges is 55.5–54.5 Ma, producing a trend of predominantly Mesozoic clasts giving way to Precambrian basement clasts upsection. Rapid source terrane unroofing is also suggested by the upsection changes in detrital zircon U-Pb ages. Detrital zircons in the upper Wind River Formation show age distributions similar to those of modern sands derived from the Wind River Range, with up to ~20% of zircons derived from the Archean basement rocks in the Wind River Range, indicating that the range was largely exhumed by ca. 53–51 Ma. The rise of these ranges by 51 Ma formed a confined valley in the northwestern part of the basin, and promoted development of a meandering fluvial system in the center of the basin. The modern paleodrainage configuration was essentially established by early Eocene time.

Carbon isotope data from paleosols and modern soil carbonates show that the soil CO₂ respiration rate during the early Eocene was higher than at present, from which a more humid Eocene paleoclimate is inferred. Atmospheric pCO₂ estimated from paleosol carbon isotope values decreased from 2050 ± 450 ppmV to 900 ± 450 ppmV in the early Eocene, consistent with results from previous studies. Oxygen isotope data from paleosol and fluvial cement carbonates show that the paleoelevation of the Wind River Basin was comparable to that of the modern Great Plains (~500 m above sea level), and that local relief between the Washakie and Wind River ranges and the basin floor was 2.3 ± 0.8 km. Up to 1 km of post-Laramide regional net uplift is required to form the present landscape in central Wyoming.

INTRODUCTION

The eastern portion of the Cordilleran orogenic belt in the western interior USA consists of the Laramide structural province, a region of Precambrian basement-involved uplifts, and intervening sedimentary basins that developed during Late Cretaceous–Eocene time (Dickinson and Snyder, 1978; Bird, 1988; DeCelles, 2004). The deformation partitioned what had been a continental-scale foreland basin that developed east of the Cordilleran thrust belt, ~1000–1500 km inland from the subduction zone. Modern regional elevation is ~1.5 km with the range summits at >4 km. Similar basement-involved uplifts in the foreland of a major Cordilleran-style orogenic system are present in the Sierras Pampeanas in South America, where flat-slab subduction is intimately associated with the region of intraforeland basin deformation (Jordan et al., 1983; Wagner et al., 2005). Although shallow subduction of the Farallon plate underneath North America is the commonly accepted tectonic cause for the Laramide deformation (e.g., Dickinson and Snyder, 1978; Bird, 1988; Saleeby, 2003; Sigloch et al., 2008; Liu et al., 2010), it remains unclear how shallow subduction would produce the individual Laramide structures and the extent to which Laramide deformation may be viewed as an eastward propagation of the greater Cordilleran strain front (Erslev, 1993; DeCelles, 2004). Critical information about the timing of individual Laramide uplifts, their paleoelevations at the time of uplift, and the temporal relationships among Laramide uplifts have yet to be documented at regional scale to address such questions.

Many previous structural, sedimentological, thermochronological, and paleoerosion studies have shown that deformation and uplift in the Laramide region is highly variable in time and space (e.g., Love, 1939; Keefer, 1957; Soister, 1968; Seeland, 1978; Gries, 1983; Winterfeld and Conard, 1983; Flemings, 1987; Dickinson et al., 1988; DeCelles et al., 1991; Gregory and Chase 1992; Cerveny and Steidtmann, 1993; Crews and Ethridge, 1993; Omar et al., 1994; Strecker, 1996; Hoy and Ridgway, 1997; Wolfe et al., 1998; Dettman and Lohmann, 2000; Crowley et al., 2002; Fricke, 2003; Peyton and Ringers, 2007; Fan and Dettman, 2009). Tectonic models explaining deformation during shallow
slab subduction include basal shear traction (Bird, 1988), thrust and back thrust connected to a master detachment in lower crust, which is possibly linked to the Sevier thrust belt (Erslev, 1993), lateral injection by mid-crustal flow from the overthickened Sevier orogenetic hinterland (McQuarrie and Chase, 2001), lithospheric buckling in response to horizontal endload (Tikoff and Maxson, 2001), and isostatic rebound by removing the ecologized Shatsky conjugate plateau on the Farallon plate (Liu et al., 2010). Post-Laramide modification of the lithosphere in the Laramide region may have played a vital role in shaping the modern landscape. Mechanisms of modification include: (1) thermal uplift due to asthenospheric upwelling by removing the subducted slab or thickened mantle lithosphere beneath the western USA (Dickinson and Snyder, 1978; Humphreys, 1995; Sonder and Jones, 1999); (2) subcontinental-scale dynamic subsideance by induced asthenospheric counterflow above the subducted slab (McMillan et al., 2002; Heller et al., 2003; McMillan et al., 2006); (3) regional uplift caused by isostatic rebound of lithosphere due to climate-driven erosion (Pelletier, 2009); and/or (4) thermal upwelling associated with the initiation of the Rio Grande Rift (Heller et al., 2003; McMillan et al., 2006). Quantitative data on paleoelevation, source-terrane unroofing and exhumation, climate, and paleogeography within a precise chronological context are required to evaluate these competing hypotheses. In this paper we report results of a multidisciplinary study of the sedimentology, sandstone petrography, detrital geochronology, and stable isotope geochemistry of the early Eocene basin fill in the Wind River Basin, central Wyoming. From these data we reconstruct the paleogeography, paleoclimate, source-terrane exhumation, tectonic setting, and basin evolution.

REGIONAL GEOLOGY

Stratigraphy and Age Control

The Wind River Basin in central Wyoming is one of the many Laramide intermontane basins formed to the east of the Sevier thrust belt. The basin is surrounded by reverse fault-bounded, basement-involved uplifts, including the Wind River Range on the southwest, the Washakie Range, Owl Creek Mountains, and Bighorn Mountains on the north, the Casper arch on the east, and the Granite Mountains on the south (Fig. 1). Strata of Late Cretaceous–Eocene age are exposed along the basin margin, and have been studied extensively in the past century (Keefer, 1965; Soister, 1968; Courdin and Hubert, 1969; Seeland, 1978; Phillips, 1983; Winterfeld and Conard, 1983; Love and Christiansen, 1985; Flemings, 1987). Among them, only Courdin and Hubert (1969) and Flemings (1987) presented modal petrographic analysis of Late Cretaceous and Paleocene strata in the western and southern basin. Flat-lying, lower Eocene, Wind River Formation occupies the center of the basin, with the greatest thickness along the east-northeastern basin margin (Keefer, 1965). In the northwestern corner of the basin, the slightly older Indian Meadows Formation was tilted and displaced on the top of the Wind River Formation by several north-striking thrust faults (Love, 1939; Keefer, 1957; Soister, 1968; Seeland, 1978; Winterfeld and Conard, 1983; Flemings, 1987). A moderately angular unconformity between the Indian Meadows Formation and underlying Mesozoic rocks has been observed along the southwestern flank of the Washakie Range (Winterfeld and Conard, 1983; this study). Our study focuses on the northwestern corner of the Wind River Basin (Dubois–East Fork area, Fremont County), where downcutting by the Wind River exposed ~700 m of the Indian Meadows Formation and Wind River Formation along the north bank, which is the thickest outcrop of the lower Eocene strata in the basin.

The ages of the Wind River and Indian Meadows formations are constrained by North American land mammal ages (GSA Data Repository Fig. DR1) (Love, 1939; Keefer, 1957; Winterfeld and Conard, 1983), whose boundaries in the early Eocene have been revised based on correlation and calibration with radiometric ages and magnetostratigraphy in the Green River Basin (Clyde et al., 1997; Robinson et al., 2004; Smith et al., 2008). The presence of Caninicus, Hyopsodus, and Diaecodexis places the Indian Meadows Formation in the early Wasatchian (Wa1–Wa3, 55.5–54.5 Ma), and the presence of Heptodon and Lambdotherium places the Wind River Formation in the late Wasatchian (Wa6–Wa7, ca. 53–51 Ma) (Clyde et al., 1997; Robinson et al., 2004; Smith et al., 2008). The youngest U-Pb ages of detrital zircons collected from fluvial sandstones in this study cluster at 63.3 ± 1.3 Ma, which is older than the mammalian age and thus provides no additional constraint on depositional age. The contact between the Indian Meadows Formation and the Wind River Formation is complex but appears conformable in the East Fork area (Winterfeld and Conard, 1983; this study).

Tectonic Setting

The Wyoming craton includes granite gneisses and minor amounts of metavolcanic-metasedimentary rocks older than 2.6–2.7 Ga (Frost and Frost, 1993; Frost et al., 2000). Paleo-proterozoic metavolcanic and metasedimentary rocks intruded by numerous plutons were accreted onto the Wyoming craton along the Cheyenne belt at ca. 1.86 Ga (Frost and Frost, 1993). Rifting of the western margin of Laurentia during Neoproterozoic–early Paleozoic time accommodated the siliciclastic sedimentary and metasedimentary rocks deposited under predominantly marine environments (Levy and Christie-Blick, 1991). During Paleozoic–early Mesozoic time, the location of modern western USA was in the Cordilleran miogeocline, and a thick succession of carbonate rocks and subduction siliciclastic rocks was deposited in a passive margin setting (Devlin and Bond, 1988). From Late Jurassic on, the backarc crustal shortening and thickening caused by the subduction of the Farallon plate produced the Cordilleran thrust belt, and a thick succession of marine and marginal marine sediments was deposited in Cordilleran foreland when the Western Interior Seaway inundation was terminated by the Late Cretaceous time (e.g., DeCelles, 2004). During ca. 80–40 Ma, Laramide deformation tectonically partitioned the foreland basin in Wyoming, Montana, and Colorado, and produced the basement-involved uplifts and intervening basins such as the Wind River Basin.
Basin analysis of the early Eocene Wind River Basin

A

Cordilleran accretion and batholiths (<0.25 Ga)
Wyoming-Hearne-Rae (>2.5 Ga)
Trans-Hudson (1.8–1.9 Ga)
Superior (>2.5 Ga)
Canada
USA
Yavapai-Mazatzal (1.6–1.8 Ga)
mid-continent (1.3–1.5 Ga)
Amarillo-Wichita (ca. 0.525 Ga)
Quachita orogen
Yucatan-Campeche (0.54–0.58 Ga)
East Mexico magmatic arc (0.23–0.29 Ga)
Cordilleran accretion and batholiths (<0.25 Ga)
Wyoming-Hearne-Rae (>2.5 Ga)
Trans-Hudson (1.8–1.9 Ga)
Superior (>2.5 Ga)
Canada
USA
Yavapai-Mazatzal (1.6–1.8 Ga)
mid-continent (1.3–1.5 Ga)
Amarillo-Wichita (ca. 0.525 Ga)
Quachita orogen
Yucatan-Campeche (0.54–0.58 Ga)
East Mexico magmatic arc (0.23–0.29 Ga)

B

Bighorn Mtns (2.8–2.95 Ga)
Wind River Range

C

Quaternary
Post-early Eocene deposits
Wind River Formation
India Meadows Formation
Mesozoic rocks
Paleozoic rocks
Precambrian basement

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Four schools of thought exist on the tectonic history of the Wind River Basin during Laramide deformation. Keefer (1965) suggested that the basin experienced three stages of subsidence in response to three uplift events in surrounding mountains, with the climax of deformation happening to the Casper Arch, southern Bighorn Mountains, and Owl Creek Mountains during the early Eocene. In contrast, Flemings (1987) concluded that the maximum rate of uplift and deformation was during Maastrichtian time and was caused by the thrusting of the Wind River Range and Granite Mountains. Moreover, of regional scale, Gries (1983) suggested that the NW-SE-oriented uplifts were generally earlier than the E-W-oriented uplifts, and Brown (1988) and Erslev (1993) argued progressive SW to NE arch development.

SEDIMENTOLOGY

Depositional environments are interpreted on sedimentological analysis basis of four measured sections, totaling ~750 m in thickness (Fig. 2). Paleocurrent directions are based on 211 measurements of imbricated clasts (10 per station) and trough cross-stratification (method I of DeCelles et al., 1983). The Eocene sedimentary rocks in the basin are described by Love (1939), Keefer (1957), Seeland (1978), Winterfeld and Conard (1983), and Flemings (1987), but lithofacies have not been described in detail. Lithofacies that were documented in the field are typical of fluvial deposits and are well understood in terms of depositional processes (summaries in Miall, 1978, 1996). Therefore, we provide a summary table of the most common lithofacies and physical processes in this study, and we focus on interpreting lithofacies assemblages and the corresponding depositional systems (Table 1).

Indian Meadows Formation: Stream-Dominated Alluvial Fan Association

Description

The Indian Meadows Formation consists of interlayered gray and yellow conglomerate, sandstone, and brick-red siltstone with paleosols, totaling 315 m in thickness (Fig. 2B). The Formation rests on an erosional unconformity on the upper Cretaceous Cody shale or the Wind River Formation in our studied area (Fig. 3A). The lower 100 m of the Formation (11–110 m of section 3DB) is divided into two 40- to 50-m-thick intervals of intercalated pebble-cobble conglomerate (Gch, Gct, Gcm, and Gcmi), medium-grained sandstone (Sh, Sm, and St), siltstone (Fm), and paleosol (P), sandwiching a 15 m interval of cobble-boulder conglomerate (Gcm and Gmm) (Fig. 3B). Units of lithofacies Gch, Gct, and Gcm (~5 m thick) are lenticular, overlie basal scour surfaces, and exhibit crude, upward-fining trends. In general, conglomerate grades rapidly upward into sandstone. Beds of lithofacies Sm, Sh, and St are composed of two or more, ~2- to 4-m-thick sandstone packages separated from each other by scour surfaces. Plane-parallel laminations of a few millimeters thick are present only in the upper ~0.5 m of many packages. The mudstone and siltstone beds (Fm) usually contain paleosol carbonate nodules.

The middle 100 m of the Indian Meadows Formation (7–110 m of 4DB and top of 2DB) is dominated by pebble-cobble conglomerate composed of lithofacies Gcm, Gcmi, Gct, and Gep. Two- to 4-m-thick beds are stacked to form lenticular bodies up to 30 m thick, which extend several kilometers laterally (Fig. 3C). These kilometer-scale lithosomes have erosional bases and generally fine upward. Siltstone lithofacies Fm forms beds 1–2 m thick between these conglomeratic bodies; although these units are massive, they lack distinctive paleosol features.

The upper 110 m of the Indian Meadows Formation (110–215 m of 4DB) is composed of intercalated lenticular pebble-cobble conglomerate (Gct, Gcm, Gmm, Gcmi, and Gcp), medium- to coarse-grained sandstone (Sm, Sh, and St) (Fig. 3D), siltstone (Fm), and paleosol (P). The sandstones contain floating granules and pebbles and generally cap the fining-upward conglomerates. Siltstone of lithofacies Fm (~2 m) is sandy and contains floating granules. Paleosol carbonate nodules are developed in most siltstone intervals, and organic-rich layers up to 1 m thick are preserved above some of the carbonate accumulation horizons. In some instances, several paleosols are stacked atop one another.

Interpretation

The assemblage of lithofacies in the Indian Meadows Formation is characteristic of alluvial fan depositional systems (Nemec and Steel, 1984; DeCelles et al., 1991; Stanstreet and McCarthy, 1993; Miall, 1996). The lower interval was deposited in the proximal alluvial fan by gravel-bed braided channels and debris flows. The thin bodies of conglomerate containing sedimentary structures were deposited as longitudinal bar deposits in small flashy, highly unsteady braided streams (Hein and Walker, 1977; Miall, 1978). The sandstones and siltstones were deposited on the tops or flanks of bars during waning flow or in slackwater ponds, such as abandoned channels (Rust, 1972; Miall, 1978). The unstratified, poorly sorted, matrix-supported conglomerates represent viscous debris flows triggered by catastrophic events (Nemec and Steel, 1984; Schultz, 1984; DeCelles et al., 1991; Stanstreet and McCarthy, 1993). The poorly sorted, clast-supported conglomerates represent high-concentration floods that were not in equilibrium with typical regime bedforms (Pierson, 1981; DeCelles et al., 1991).
TABLE 1. LITHOFACIES AND INTERPRETATIONS USED IN THIS STUDY*  

<table>
<thead>
<tr>
<th>Lithofacies Code</th>
<th>Description</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gmm</td>
<td>Massive, matrix-supported pebble to boulder conglomerate, poorly sorted, disorganized, unstratified</td>
<td>Deposition by cohesive mud-matrix debris flows</td>
</tr>
<tr>
<td>Gcm</td>
<td>Pebble to boulder conglomerate, poorly sorted, clast-supported, unstratified</td>
<td>Deposition from sheet floods and clast-rich debris flows</td>
</tr>
<tr>
<td>Gcmi</td>
<td>Pebble to boulder conglomerate, moderately sorted, clast-supported, stratified, imbricated</td>
<td>Deposition by traction currents in unsteady fluvial flows</td>
</tr>
<tr>
<td>Gct</td>
<td>Pebble to cobble conglomerate, well sorted, clast-supported, trenched cross-stratified</td>
<td>Deposition by large gravely ripples under traction flows in relatively deep, stable fluvial channels</td>
</tr>
<tr>
<td>Gchi</td>
<td>Pebble to cobble conglomerate, well sorted, clast-supported, trenched cross-stratified, imbricated (long axis transverse to paleoflow)</td>
<td>Deposition from shallow traction currents in longitudinal bars and gravel sheets</td>
</tr>
<tr>
<td>Gch</td>
<td>Pebble to cobble conglomerate, well sorted, clast-supported, horizontally stratified</td>
<td>Deposition from shallow traction currents in longitudinal bars and gravel sheets</td>
</tr>
<tr>
<td>Gcp</td>
<td>Pebble to cobble conglomerate, well sorted, clast-supported, planar cross-stratified</td>
<td>Deposition by large straight-crested gravely ripples under traction flows in shallow fluvial channels</td>
</tr>
<tr>
<td>Sm</td>
<td>Massive very coarse- to medium-grained sandstone, sometimes contains gravel</td>
<td>Deposition from small gravity flow</td>
</tr>
<tr>
<td>Sr</td>
<td>Fine- to medium-grained sandstone with small, asymmetric ripples</td>
<td>Migration of small 2D and 3D ripples under weak unidirectional flows in shallow channels</td>
</tr>
<tr>
<td>St</td>
<td>Medium- to coarse-grained sandstone with trough cross-stratification</td>
<td>Migration of large 3D ripples (dunes) under moderately powerful unidirectional flows in large channels</td>
</tr>
<tr>
<td>Sh</td>
<td>Fine- to medium-grained sandstone, horizontally stratified</td>
<td>Upper plane bed conditions, or flash floods under unidirectional flows, either strong or very shallow</td>
</tr>
<tr>
<td>Fr</td>
<td>Siltstone, horizontally laminated, or small ripples</td>
<td>Deposition from suspension or weak traction current in overbank area</td>
</tr>
<tr>
<td>Fm</td>
<td>Massive siltstone, sometimes bioturbated</td>
<td>Deposition from suspension in ponds and floodplain</td>
</tr>
<tr>
<td>P</td>
<td>Massive siltstone with carbonate nodules</td>
<td>Calcic paleosol</td>
</tr>
</tbody>
</table>

*Modified from Miall (1978) and DeCelles et al. (1991).

The thick conglomerate beds in the middle interval of the Indian Meadows Formation are typical of shallow gravel-bed braided river channels, with cross-stratification interpreted as bar accretion sets, and the upward-fining trends reflecting a progressive decrease in flow strength (Hein and Walker, 1977; Miall, 1996). The absence of debris flow facies and extensive lateral distribution of this interval indicates the depositional environment was an extensive gravelly braided plain (Rust and Gibling, 1990). The upper interval was deposited in a stream-dominated alluvial fan environment, with the sandstone and siltstone containing floating pebbles deposited in sandy mudflows that spilled out of channels onto the preexisting ground surface as sheet floods (Pierson, 1981). The siltstones contain bioturbated paleosols and sparse carbonates suggest that the floodplain of the braided rivers was vegetated and the paleosols were well developed when the mean annual precipitation was high, and the climate strongly seasonal (Retallack, 2005).

The paleocurrent direction varies between eastward and southwestward, but the most prominent direction was eastward (Fig. 2C). This suggests that the sediments were mainly derived from the Wind River Range to southwest, with local input from the Washakie Range and/or western Owl Creek Mountains to north (Figs. 1B and 1C).

Overall, the depositional environment in the northwestern Wind River Basin evolved from alluvial fan into low-sinuosity fluvial, with paleocurrent directions changing from southwestward to mostly eastward during 55.5–51 Ma. This evolution suggests that surface slope between the upland sediment source terranes and the basin decreased through time (Stanistreet and McCarthy, 1993; Miall, 1996).

SANDSTONE PETROGRAPHY AND PROVENANCE

Methods and Description

Standard petrographic thin sections were made from 16 medium-grained sandstone samples that were collected while measuring stratigraphic sections. The thin sections were stained for potassium and calcium feldspar and point-counted (450 points per slide) using a modified Gazzi-Dickinson method, in which crystals larger than silt-sized within lithic fragments are counted as monocrystalline grains (Ingersoll et al., 1984). The point-counting parameters are listed in Table DR1 (see footnote 1), and modal data are given in Table DR2 (see footnote 1). The modal sandstone analyses are augmented by 11 clast-counts (at least 100 clasts per station).

Monocrystalline quartz is the primary constituent of all samples. Feldspar mainly consists of potassium feldspar and plagioclase, with a K/P ratio that varies between 0.6 and 4.7. Lithic

The assemblage of lithofacies in the Wind River Formation is characteristic of low-sinuosity, gravel-bed braided river depositional systems (Hein and Walker, 1977; Miall, 1978, 1996). The conglomerates represent braided channel deposits, and the sandstones were deposited on the tops or flanks of bars or in abandoned channels. The stacked lenticular thin beds of horizontally stratified and cross-stratified sandstone are interpreted as crevasse-splay deposits (Miall, 1978, 1996). The lack of floating pebbles in siltstones suggests the lack of sandy mudflows (Pierson, 1981). Thick paleosol carbonate horizons suggest that the floodplain of the braided rivers was vegetated and the paleosols are well developed when the mean annual precipitation was high, and the climate strongly seasonal (Retallack, 2005). The paleocurrent direction varies between eastward and southwestward, but the most prominent direction was eastward (Fig. 2C). This suggests that the sediments were mainly derived from the Wind River Range to southwest, with local input from the Washakie Range and/or western Owl Creek Mountains to north (Figs. 1B and 1C).
grain populations are dominated by micritic carbonate. Volcanic grains constitute <5% of total lithic grains. Other lithic grains include phyllite, siltstone, and quartzite derived from sedimentary and metasedimentary strata. Biotite, muscovite, pyroxene, olivine, glauconite, and chlorite are the most common accessory minerals. The Indian Meadows and Wind River formations sandstones have average modal compositions of Qm:F:Lt = 57:10:33, Qt:F:L = 77:11:13; and Qm:F:Lt = 41:26:33, Qt:F:L = 56:26:18, respectively. The Indian Meadows Formation contains greater amounts of quartz, and lesser amounts of feldspar than the Wind River Formation (Fig. 4). All of the Eocene sandstones that we studied are dominated by calcite cements (Fig. 5).

Clasts of dolostone, limestone, and sandstone dominate the conglomerates in the Indian Meadows Formation. Carbonate clasts with brachiopod fossils are abundant in the lower interval of the Indian Meadows Formation. The sandstone clasts are quartzose and have distinct red color. Granite gneiss clasts first appear in the middle interval of the Indian Meadows Formation, and the content increases upsection, reaching ~20% at the top of the Indian Meadows Formation. Granite (~45%) and carbonate (~40%) clasts dominate the conglomerates in the Wind River Formation.

**DETRITAL ZIRCON U-Pb GEOCHRONOLOGY**

**Samples and Methods**

Six samples of medium- to coarse-grained sandstone from the Indian Meadows Formation (3DB69, 4DB4.5, 4DB191, and 2DB273) and Wind River Formation (2DB2 and 2DB257), one sample of a granite cobble (2DB2cobble), and two samples of modern river sands (East...
Fork (EF), Little Popo Agie River (LP)) were collected and processed by standard methods for separating zircons (Gehrels et al., 2000). The Little Popo Agie River and East Fork rise in Precambrian crystalline basement of the Wind River Range and Washakie Range, respectively (Fig. 1B). The former cuts an ~2-km-thick Paleozoic and Mesozoic sedimentary section on the northeastern flank of the Wind River Range, whereas no Paleozoic and Mesozoic sedimentary strata are exposed along the East Fork. Zircon grains in sample EF are euhedral, and ~200 μm long, whereas those in sample LP are similar to those from Eocene sandstones, subhedral to anhedral, and mostly 50–100 μm in diameter (Fig. DR2 [see footnote 1]).

About 100 individual zircon grains were analyzed for the early Eocene sandstone samples and sample LP. All 19 zircons extracted from sample EF and 23 zircons from sample 2DB2cobble were analyzed. U-Pb geochronology of zircons was conducted by laser-ablation–multi-collector–inductively coupled plasma–mass spectrometry (LA-MC-ICP-MS) at the University of Arizona LaserChron Center. Details of analytical procedures and data processes are described in Gehrels et al. (2008). The ages presented are ^{206}Pb/^{238}U ages for grains <1000 Ma and ^{206}Pb/^{208}Pb ages for grains with ^{206}Pb/^{238}U ages >1000 Ma (age-probability plots in Fig. 6; concordia diagrams in Fig. DR3 [see footnote 1]). Analyses that yielded isotopic data of acceptable discordance, in-run fractionation, and precision are shown in Table DR3 (see footnote 1). Nine of the 19 zircon grains from sample EF show concordant ages. There is no systematic correlation between shape, color, angularity, and U-Pb age for zircons in the Eocene sandstones and sample LP.

Results

The youngest detrital zircons recovered from the Eocene basin fill are in the 60 to 80 Ma range, which is older than the depositional age derived from North American land mammal ages. Ages of pre–80 Ma zircons generally fall within the following intervals: 80–220 Ma, 230–290 Ma, 330–760 Ma, 1.0–1.3 Ga, 1.3–1.5 Ga, 1.6–1.8 Ga, and 2.5–3.0 Ga. Zircons of Proterozoic age constitute the largest population on age-probability diagrams. Sample 2DB2cobble has a mean U-Pb zircon crystallization age of 2.61 Ga (Fig. 7). A single zircon grain in sample LP produced an age of 47.7 ± 2.1 Ma, but the other age groups are similar to those in the Eocene samples. The nine detrital zircon grains with concordant ages in sample LP cluster at 2.7–3.2 Ga. The interpretation of detrital zircon ages in the western USA is relatively well understood (Armstrong and Ward, 1993; Gehrels et al., 1995; Roback and Walker, 1995; Stewart et al., 2001; Torres et al., 1999; Gehrels, 2000; Gehrels et al., 2000; DeGraaff-Surpless et al., 2002; Dickinson and Gehrels, 2003; Link et al., 2005; Dickinson and Gehrels, 2009). Here, we summarize the sources of specific age groups in Table 2. The major source terranes are presented in Figure 1A.

Interpretation

The euhedral morphology of zircon grains suggests sample EF was derived from weathered granite in the Washakie Range and/or the Owl Creek Mountains. The zircon U-Pb ages are consistent with the crystallization age of the granodiorite gneiss in west Owl Creek Mountains (ca. 2.8 Ga) (Kirkwood, 2000). The youngest zircon in sample LP is ca. 48 Ma; this grain was most likely derived from the Absaroka volcanic field (Armstrong and Ward, 1993). Approximately 20% of the detrital zircons analyzed in sample LP are in the 2.5 to 2.7 Ga age range, which is consistent with zircon U-Pb ages of the granitic plutons in the Wind River Range (2.55–2.67 Ga) (Frost et al., 2000). The anhedral morphology and small grain size suggest most of the zircon grains in sample LP were recycled, presumably from Mesozoic sandstones. This is supported by the fact that Mesozoic sandstones in western USA have similar age spectra, including: (1) zircons from the Cordilleran magmatic arc in California and Nevada (80–220 Ma) and
Figure 6. U-Pb age-probability diagrams of detrital zircons for samples in this study. The curves represent a sum of the probability distributions of all analyses from a sample, normalized so that the areas beneath the probability curves are equal for all samples. Thick dashed line represents 2.61 Ga, which is marked as the reference of Archean zircons from the Wind River Range (see Fig. 7). Sources of different zircon age peaks in this study are summarized in Table 2.
Permian–Triassic magmatic arc sources in eastern Mexico (230–280 Ma); (2) recycled zircons from the Devonian–Mississippian Antler orogenic belt in Nevada and Idaho (330–450 Ma); (3) zircons derived from the Appalachian orogenic belt (360–760 Ma), and Grenville terrane (1.0–1.3 Ga) in the eastern USA; and (4) recycled zircons from the Yavapai-Mazatzal orogen (1.6–1.8 Ga) (Roback and Walker, 1995; Dickinson and Gehrels, 2003, 2009).

Detrital zircon geochronology of the modern river sands sheds light on the Eocene detrital zircon age spectra in the following aspects: (1) zircon grains recycled from Mesozoic sandstones with mostly Grenville and Yavapai-Mazatzal ages, dominate the zircon population in fluvial basinal sediments; (2) basement exposures along the modern basin margin contribute only ~20% of the total zircon grains to basinal sediment; and (3) zircons from the Washakie Range (2.7–3.1 Ga) are older than zircons from the Wind River Range (ca. 2.6 Ga).

The evolution of the detrital zircon age spectra in the Indian Meadows Formation shows a rapid source terrane unroofing in 55.4–54.5 Ma (Fig. 6). The detrital age spectrum of the lower part of the formation (3DB69 and 4DB4.5) is similar to that of the Jurassic eolianite in Utah and Wyoming, with abundant Grenville (ca. 1.0–1.3 Ga) and lesser amounts of Yavapai-Mazatzal (1.6–1.8 Ga) age grains (Dickinson and Gehrels, 2003, 2009). The proportion of grains derived from the Cordilleran magmatic arc increases in the middle formation (sample 2DB273). This change corresponds to the inferred increase of fluvial discharge in the alluvial fan association. The age spectrum of the upper formation (sample 4DB191) is dominated by grains of Yavapai-Mazatzal age, reflecting a recycled source in Paleozoic strata (Gehrels et al., 1995; Stewart et al., 2001) when the Washakie and/or western Owl Creek ranges were unroofed.

The detrital zircon age spectra in the Wind River Formation show sediment shed from the Wind River Range dominated the basal sediment in 53–51 Ma. The ca. 2.61 Ga zircon age from the granite cobbled from the lower Wind River Formation matches the age of basement granite in the Wind River Range (Frost et al., 2000). Abundant granite clasts and Archean zircons (ca. 2.6 Ga) in the lower Wind River Formation (sample 2DB2) suggest that Wind River Range basement rocks were a major source by early Eocene time. The detrital zircon age spectrum changes into the shape of modern sand sample LP at the top of the Wind River Formation (sample 2DB257), with predominant Yavapai-Mazatzal ages, a large component of Mesozoic ages indicating derivation from the Cordillera magmatic arc, and ~20% Archean ages with sources in the Wind River Range basement rocks.

In summary, detrital zircon geochronology of lower Eocene sandstones in the Wind River Basin shows: (1) the source terrane of the Indian Meadows Formation (ca. 55.5–54.5 Ma) was most likely the Washakie Range and/or Owl Creek Mountains, consistent with southwestward paleoflow directions. Sediment derived from these sources records an unroofing sequence from upper Mesozoic sandstone to lower Paleozoic strata; (2) grains derived from the Wind River Range began to enter the depositional system by ca. 53–51 Ma; and (3) the basement of the Wind River Range was eroded and exposed to its present extent by 51 Ma.

### STABLE ISOTOPE GEOCHEMISTRY

**Methods**

The basic approach of this study and the analytical methods for carbonate isotope analysis follow that of DeCelles et al. (2007). Modern soil carbonate formed under cobbles clasts from at least 50 cm depth in active soil horizons was collected. Paleosol carbonate nodules were collected from the middle of paleosol beds at least 50 cm below the paleosurface (Fig. 8). We also analyzed the δ13C values of eight dominant plant species in the study area collected during the growing season. Organic matter δ13C was measured using a Costech EA combustion unit connected to a Finnigan Delta Plus XL mass spectrometer via a ConFlO III split interface. The precision is better than ±0.1% (1σ). All the isotope values of carbonate and organic matter are reported relative to PeeDee belemnite (PDB), and that of water relative to standard mean ocean water (VSMOW).
The δ13C values of all marine and nonmarine carbonate in the Wind River Basin range from –17.6‰ to –1.5‰ (Table DR4 [see footnote 1]; Fig. 9). Recycled marine limestone clasts and coral fossils have the highest δ13C values, from –17.6‰ to –11.5‰, whereas early Eocene fluvial cements have the lowest values, from –17.6‰ to –11.5‰. The δ13C values of early Eocene paleosol nodules are lower than the values of micrite in the paleosol nodules (Fig. 9). This suggests that the calcite veins probably formed by fluids at temperatures higher than at the Eocene soil surface. If paleosol nodule micrite underwent diagenetic resetting during the formation of secondary sparry veins, the isotopic values of micrite would be similar to the values of secondary veins. Third, the δ18O and δ13C values of recycled marine limestone clasts and fossil corals in conglomerates immediately adjacent to paleosols and sandstones are similar to those from unaltered Mesozoic–Paleozoic marine carbonate (Veizer et al., 1999) but significantly higher than the values from paleosol and fluvial carbonate cement. The δ18O and δ13C values of soil nodules, secondary veins, fluvial cement, and marine limestone are distinct from early cementation occurs prior to significant compaction (Quade and Roe, 1999, and references therein; Sanyal et al., 2005). Second, the δ18O values of a few sparry calcite veins in paleosol nodules are lower than the values of micrite in the paleosol nodules (Fig. 9). This suggests that the calcite veins probably formed by fluids at temperatures higher than at the Eocene soil surface. If paleosol nodule micrite underwent diagenetic resetting during the formation of secondary sparry veins, the isotopic values of micrite would be similar to the values of secondary veins. Third, the δ18O and δ13C values of recycled marine limestone clasts and fossil corals in conglomerates immediately adjacent to paleosols and sandstones are similar to those from unaltered Mesozoic–Paleozoic marine carbonate (Veizer et al., 1999) but significantly higher than the values from paleosol and fluvial carbonate cement. The δ18O and δ13C values of soil nodules, secondary veins, fluvial cement, and marine limestone are distinct from each other, and form clusters in the δ13C versus δ18O diagram (Fig. 9). Diagenesis would tend to homogenize the isotopic values of marine limestones, micrites, and secondary spars. The large variation of fluvial cement δ13C values may be due to contamination during sampling by marine limestone detritus, which have relatively positive values (Figs. 9 and 10).

**Oxygen Isotopes and Paleoaltimetry**

The δ18O value of ancient local meteoric water can be calculated from paleosol δ18O values and constrained temperature. Pedogenic
calcite is thought to form in isotopic equilibrium with soil water during the warm season (Quade et al., 1989; Cerling and Quade, 1993; Quade et al., 2007; Breeker et al., 2009). The temperature of the warm season is usually between the mean annual temperature and maximum summer temperature. Soil water derives from rainfall that may undergo evaporative enrichment in δ18O, especially in arid settings. We assume the summer temperature in Wyoming in early Eocene was ~30 °C based on an early middle Eocene temperature reconstruction in France, which has a similar mean annual temperature as Wyoming (Andreasen and Schmitz, 1996). The δ18O values of early Eocene precipitation can then be estimated by using soil nodule formation temperatures of 18.6–30 °C and standard calcite fractionation factors (Kim and O’Neil, 1997). The calculated δ18O values of early Eocene precipitation in the Bighorn and Wind River basins are −6.8‰ ± 1.9‰ (VSMOW), which are consistent with the values derived from the δ18O values of mammal tooth enamel in the early Eocene Bighorn Basin and Powder River Basin (Fricke, 2003) and the values derived from freshwater bivalve fossils in the lower Eocene in the Powder River Basin (Fan and Dettman, 2009). For comparison with modern water, 1‰ is added to the δ18O value of early Eocene precipitation to account for the lower seawater δ18O value in the early Eocene (Zachos et al., 1994).

The δ18O value of groundwater can be calculated from fluvial cement δ18O values and constrained temperature. Early diagenetic cements in fluvial sandstone formed in equilibrium with shallow groundwater (Quade and Roe, 1999; Mack et al., 2000). The deposition temperature of the early cements should be close to mean annual temperature since it formed near the surface (18.6 ± 2.4 °C). The calculated δ18O value of groundwater is −16.6‰ ± 0.5‰ (VSMOW), using the most negative δ18O value of fluvial cement (−17.6‰, PDB). This value (−16.6‰) is comparable with the unweighted mean δ18O value of modern precipitation in the south flank of the Wind River Range (−15.2‰, Pinedale, Wyoming, elevation of 2.4 km) after adding 1‰ for the lower early Eocene seawater δ18O value (Zachos et al., 1994) (http://www.uaa.alaska.edu/envir/envisnp). The large difference between the δ18O values of groundwater and local precipitation in the early Eocene Wind River Basin suggests the groundwater was not sourced in local basin precipitation, but rather from the surrounding highlands.

At present, rainfall in the high elevations of the Rocky Mountains is dominated by winter precipitation from the Pacific Ocean and the Arctic Ocean. The low elevation regions in the Laramide province likely received abundant summer precipitation from the Gulf of Mexico and small amounts of winter precipitation from the Arctic Ocean (Bryson and Hare, 1974; Dutton et al., 2005). Today, the elevation along a west-east transect through the Rocky Mountains, Laramide province, and Great Plains decreases from 3–4 km to 0.5 km, which strongly influences the δ18O value of modern precipitation and the local isotopic lapse rate (Fig. 11). By early Eocene time, crustal shortening in the Sevier had caused significant crustal thickening, and formed an elevated hinterland plateau to the west of the Laramide province (DeCelles, 2004). Therefore the paleotopography of Wyoming in early Eocene time was probably similar to present, with the main vapor source from the ancestral Gulf of Mexico. This justifies the comparison of the δ18O values of the early Eocene precipitation with the modern isotopic patterns.

Warmer global temperature in early Eocene time could affect the relationship between the δ18O value of precipitation and local temperature. Therefore, we compare the δ18O values of early Eocene precipitation in the Wind River Basin with modern summer precipitation isotopic gradient in the region (Fig. 11). The δ18O values of early Eocene precipitation (−5.8‰ ± 1.9‰ after correction) in the Wind River and Bighorn basins are similar to the values of the

![Figure 10. Simplified stratigraphy, carbon and oxygen isotope values of paleosol and fluvial cement, percentage of granite clasts, Archean zircons, and feldspar through section. Vertically ruled areas are depositional lacunae. PDB—PeeDee belemnite.](image)
modern summer precipitation at the same latitude of the Great Plains (~5.5‰, Mead, Nebraska, elevation of 352 m) (Harvey, 2001), but significantly higher than the values of modern summer precipitation in the Laramide province (~12.6‰, Butte, Montana, elevation of 1653 m) (Gammons et al., 2006), suggesting the paleo-elevation of the early Eocene Wind River Basin and Bighorn Basin was low (~500 masl). The low basinal elevation of the Wind River Basin was stable in early Eocene because there is no systematic variation of the δ18O value of paleosols through the stratigraphic section (Fig. 10). The difference (9.8‰ ±2.0‰) between the δ18O values of early Eocene river water and basinal rainfall suggests that the elevation difference between the surrounding Laramide ranges and basin floor was 3.4 ± 0.7 km, based on the isotopic lapse rate of -2.9‰/km derived from modern precipitation in the USA (Dutton et al., 2005). Such an elevation difference is higher than the modern difference of 1.5–2.5 km. The high estimated elevations of the Laramide ranges (~4 km) from our study are consistent with the elevation estimates derived in previous paleobotanical studies for the early Eocene (Gregory and Chase, 1992; Wolfe et al., 1998).

Carbon Isotopes, Paleoclimate, and pCO2

The δ13C value of soil carbonate is determined by the relative proportion of C3 to C4 plants in moist climate, and by the extent of local plant cover in arid climate where soil respiration rates are low (Quade et al., 1989; Cerling and Quade, 1993; Breeker et al., 2009). Modern climate in the Wind River Basin is semiarid, with the mean annual precipitation of ~200 mm. Our analyses show that the most common grasses and the dominant plant, sagebrush, in the sparsely vegetated basin are C3, with the δ13C values of ~24.5‰ ± 1.0‰. Soil carbonate formed in equilibrium with CO2 respired from pure C3 plants will have a δ13C value of ~12‰ to ~10‰ (Cerling and Quade, 1993). The relatively high δ13C values of modern soil carbonate in the Wind River Basin (~2.4‰ ± 2.1‰) clearly suggests mixing of atmospheric CO2 with plant-derived CO2 owing to low soil respiration rates in the local semiarid climate (Fig. 9).

The δ13C values of early Eocene paleosols (~7.0‰ ± 2.0‰) in the Wind River Basin are significantly lower than the values of modern soil carbonate (Fig. 9), which suggests a wetter climate during the Eocene. This is supported by several other lines of evidence: (1) calcisols are well developed in the floodplain of braided rivers in early Eocene Wind River Basin. The thick (>50 cm) reddish argillic horizons that typify Wind River Formation paleosols point to a relatively moist setting; and (2) abundant leaf fossils of mixed deciduous and evergreen plants in the Wind River Formation show high mean annual temperature up to 21 °C and paleoprecipitation up to 2000 mm/yr (Hickey and Hodges, 1975; Wilf et al., 1998).

However, the δ13C values of early Eocene paleosols are higher than the values of soil carbonate formed in wet settings today, in the presence of C3 biomass. This can be explained by the presence of C4 plants or higher pCO2 during early Eocene time. There is no robust evidence for the existence of C4 plants in early Eocene (Cerling and Quade, 1993). This justifies the use of δ13C values of paleosols to estimate ancient atmospheric pCO2 levels based on a diffusion-based model developed by Cerling (1992), under the assumption of elevated soil respiration rates expected in moist soil-forming conditions. δ13C values of soil carbonate are determined by the δ13C value of soil CO2, which is a mixture of soil-respired CO2 and atmospheric CO2. Therefore, the δ13C value of soil CO2 is a function of the concentration (pCO2) and δ13C value of CO2 in the atmosphere, and the concentration and δ13C value of soil-respired CO2. The δ13C value of ancient soil CO2 can be calculated from the measured δ13C value of paleosols, using the temperature-dependent fractionation equation (Romanek et al., 1992). The concentration of soil-respired CO2 is a function of soil respiration rate and soil depth. Carbonate nodules were collected at least ~50 cm below the paleosurface (the top of the paleosol profile). In our reconstruction we assume: (1) soil respiration rates ≥4 mmol/m²/hr (Cerling, 1992; Ekart et al., 1999); (2) the δ13C value of CO2 in the atmosphere is assumed to be near pre-industrial values of ~6.5‰, and the values of plant-derived CO2 to be ~24.5‰; and (3) an exponential CO2 production function (Quade et al., 2007).

Using the average value of the three most positive δ13C values of paleosols in the lower Indian Meadows Formation, atmospheric pCO2 at ca. 55 Ma is estimated to have been 2050 ± 450 ppmV (based on 1σ of the δ13C value of the paleosol carbonate). Using the average value of the more negative δ13C values in the upper Indian
Meadows and Wind River formations, atmospheric pCO₂ from 54 to 51 Ma is estimated to have been 900 ± 450 ppmV. The calculated high pCO₂ values and declining early Eocene trend, based on paleosols in central Wyoming, are consistent with previous estimates of Eichholtz et al. (1999) using the same method, and by Pearson and Palmer (2000) using boron isotope ratios of planktonic foraminifer shells.

REGIONAL PALEOGEOGRAPHY

Our reconstruction of the paleogeography in the northwestern Wind River Basin during the early Eocene is divided into three temporal stages constrained by the relative stratigraphic position of particular lithofacies associations, the paleoelevations of the surrounding ranges, the change in sedimentary provenance recorded by detrital zircon age spectra, and absolute ages based on fossil assemblages in the basin (Fig. 12).

The lower Indian Meadows Formation was deposited unconformably on upper Cretaceous Cody shale in the East Fork area. The depositional hiatus formed at least partly by folding in front of the Washakie Range (Winterfeld and Conard, 1983). Deposition evolved from small braided rivers into debris flows on proximal alluvial fans along the range front. Southward-flowing debris flows delivered mainly Mesozoic and Paleozoic detritus. The gravitational force associated with steep slope is important to the initiation of debris flows (Stanistreet and McCarthy, 1993; Coussot and Meunier, 1996). Therefore, we infer that the Washakie Range to the north of the region was high during the early Eocene (Fig. 12A), which is also supported by stable isotope paleoaltimetry (~4 km).

Deposition in the upper Indian Meadows Formation was dominated by shallow braided rivers on alluvial fans (Fig. 12B). Southward flow delivered late Paleozoic clasts from the Washakie Range into the basin, and the Precambrian basement core in the Washakie Range was gradually exposed. The change from debris flows to braided rivers suggests the slope between the mountain range and basin floor decreased (Stanistreet and McCarthy, 1993). Such a change of slope may have been caused by the rapid erosion of the mountain range and accumulation of sediment in wet climate.

Braided rivers flowing east dominate the sedimentary environment of the Wind River Formation (Fig. 12C). We argue that the change of paleoflow direction was caused by the development of a northeastward paleoslope along the Wind River Range during or after the deposition of the Indian Meadows Formation, and before deposition of the Wind River Formation. Paleoslopes along both the Wind River and Washakie ranges formed a confined valley during the early Eocene, and low-sinuosity rivers were developed in the narrow basin when discharge and sediment yield were high. The dominant northeasterward paleocurrent direction and granitic cobbles derived from the Wind River Range suggest that the range became the major source terrain of sediment in the northwest part of the basin from ca. 53 to 51 Ma. The abundance of basement granite clasts, feldspar, and Archean zircons suggests that the basement core of the Wind River Range had been exposed by ca. 53 Ma (Fig. 10). Low-sinuosity rivers that developed upstream could have merged into a large east-flowing meandering river in the basin (Soister, 1968; Seeland, 1978). Therefore, a paleodrainage pattern similar to the present one was developed by late early Eocene time in the northwest Wind River Basin. The local relief based on oxygen isotope ratios was 2.3 ± 0.8 km, which is also comparable to modern relief in the region, but with lower basin floor.

IMPLICATIONS FOR TECTONICS

Rapid Late Paleocene–Early Eocene Uplift of Basement-Involved Ranges

Both the Wind River and Washakie ranges had an elevation contrast of 2.3 ± 0.8 km relative to the basin by late Paleocene–early Eocene time, and the basin floor stood at ~0.5 km (asl). Rapid late Paleocene–early Eocene tectonic uplift and growth of surface topography throughout western Wyoming are supported by evidence for synorogenic sedimentation, reflection seismic records, and thermochronology. Alluvial fan conglomerates of late Paleocene–early Eocene age, indicating intense unroofing of Laramide source terranes, also occurred along the east flank of the Beartooth Range, east of the Bighorn Range, north of the Uinta Mountains, and on the south flank of the Wind River Range (Gries, 1983; DeCelles et al., 1991; Crews and Ethridge, 1993; Hoy and Ridgway, 1997). Seismic and borehole records from the south flank of the Wind River Range show that
Precambrian rocks are in thrust contact above the lower Eocene sedimentary rocks (Gries, 1983, and references therein). Thermochronological studies of the Beartooth Range and Wind River Range also show that of the mountains exhumed 4–8 km in late Paleocene–early Eocene time (Omar et al., 1994; Peyton and Reiners, 2007). This exhumation amount is significantly higher than the thickness of Phanerozoic sedimentary cover on the Precambrian cores.

Uplift of the Wind River Range may have started in Late Cretaceous as a paleosol developed on the base of the Paleocene deposits on the east flank of the Wind River Range (Gries, 1983). Uplift during Late Cretaceous would be consistent with data from numerous other Laramide uplifts and basins. For example, the Hoback Basin to the west of the Washakie and Wind River ranges contains thick uppermost Cretaceous and Paleocene sediments with abundant conglomerates (Love, 1973); and the upper Cretaceous–lower Paleocene Bearhead Conglomerate and Maastrichtian Sphinx Conglomerate in southwestern Montana indicate uplifting of the Blacktail-Snowcrest and Madison-Gravelly arches, respectively (Haley, 1986; DeCelles et al., 1987). It is not clear how such evidence of Late Cretaceous tectonism of the Laramide ranges was expressed in terms of paleoelevations, but the abundance of coarse-grained alluvial fan deposits containing proximal sediment-gravity flows, as well as provenance studies, indicate substantial topographic relief (Graham et al., 1986; DeCelles et al., 1991). Large-scale tectonic exhumation and unroofing of the Laramide ranges in western Wyoming during late Paleocene–early Eocene time could have been enhanced by the relatively humid paleoclimate. Studies in the Himalaya and Washington Cascades have shown that high precipitation is coupled with rapid erosion and sediment detachment rate (Reiners et al., 2003; Thiede et al., 2005).

Post-Early Eocene Regional Uplift

Currently the Laramide basins in Wyoming have an average elevation of ~1.5 km, roughly 1 km higher than their elevations during the early Eocene, based on oxygen isotope ratios of basinal precipitation recorded in paleosol nodules. Therefore, the Laramide basin floors in Wyoming must have reached their present elevation after early Eocene time.

Three mechanisms are proposed to explain the surface uplift of Laramide basins after early Eocene: (1) removal of Farallon slab or thickened mantle lithosphere in western USA in middle Cenozoic time (Dickinson and Snyder, 1978; Humphreys, 1995; Sonder and Jones, 1999; Liu et al., 2010), (2) late Cenozoic isostatic rebound of lithosphere due to climate-controlled or enhanced sediment erosion (Pelletier, 2009), and (3) late Cenozoic mantle thermal uplift caused by the thermal upwelling associated with the initiation of the Rio Grande Rift (Heller et al., 2003; McMillan et al., 2006).

Southward migration of post-Laramide magmatism in the northern Basin and Range is argued as the evidence for southward removal of Farallon slab or mantle lithosphere (Humphreys, 1995; Sonder and Jones, 1999). The resulting asthenospheric upwelling could have caused regional uplift. Post-Laramide magmatism occurred in northwestern Wyoming and Idaho at ca. 50 Ma (Armstrong and Ward, 1993), which predicts the timing of regional uplift in the Laramide province in Wyoming. However, there is no robust evidence suggesting high basinal elevation before early Eocene (Wolfe et al., 1998; Sjostrom et al., 2006; this study). Therefore it is unlikely that the removal of Farallon slab caused the post-early Eocene regional elevation gain in the Laramide foreland.

Thermo-geochronological and sedimentological data suggest an episode of Miocene exhumation in the Laramide ranges, which could have caused regional surface uplift (Cerveny and Steidtmann, 1993; Omar et al., 1994; Streck, 1996; Crowley et al., 2002; Peyton and Reiners, 2007). A recent reconstruction of post-Laramide basin fill also shows that more than 1 km of incision occurred in the Laramide basins and the western Great Plains since late Miocene time (McMillan et al., 2006). Therefore, the Miocene uplift due to climate-enhanced erosion and/or mantle dynamic uplift most likely formed the present landscape in central Wyoming (Heller et al., 2003; McMillan et al., 2006). However, more paleoelevation data covering Cenozoic time are required to evaluate the amount of regional uplift during the late Cenozoic. Our research does not preclude dynamic uplift between early Eocene and late Miocene time.

CONCLUSIONS

This multidisciplinary study of the lower Eocene strata in the northwestern corner of the Wind River Basin improved our understanding of the timing and process of basin evolution and source terrane unroofing, changes in paleoelevation of basin floor and surrounding Laramide uplifts, and paleoclimate during Laramide deformation.

(1) Petrographic, detrital zircon U-Pb geochronology, and paleocurrent analyses show the Washakie Range and western Owl Creek Mountains to the north of the basin experienced rapid unroofing at 55.5–54.5 Ma, and the Wind River Range to the southwest became the dominant sediment source terrane by ca. 53–51 Ma. Rapid tectonic uplift of the two mountains in the late Paleocene–early Eocene formed a confined valley, and a paleodrainage similar to the present with rivers flowing from both ranges was formed in the northwestern Wind River Basin by 51 Ma. The sedimentary environment evolved from alluvial fan into low-sinuosity, gravel-bed braided rivers, and paleocurrents changed from south-westward into mostly eastward.

(2) Detrital zircon grains equivalent to the depositional age are absent in early Eocene sediment. Recycled detrital zircon grains derived from the Grenville and Yavapai-Mazatzal orogens are present in Mesozoic and Paleozoic sedimentary rocks exposed in the Laramide ranges, and dominate the zircon populations of both modern river sand and early Eocene sandstones in the Wind River Basin. Presently exposed crystalline Archean basement in the Wind River Range contributes ~20% of the total zircon grains to basinal sediments. Zircons of different ages from Precambrian basement rocks exposed in Laramide ranges are a potentially useful tracer of sedimentary provenance.

(3) A more humid climate in the early Eocene Wind River Basin is inferred from the high soil CO₂ respiration rate compared to today. Calcisol soils were extensively developed in the floodplain deposits. Atmosphere pCO₂ estimated from paleosol δ13C values decrease from 2050 ± 450 ppmV to 900 ± 450 ppmV during the early Eocene, consistent with previous reconstructions of pCO₂ for this period.

(4) The elevation of the Wind River Basin was comparable with the modern Great Plains, on the order of ~500 m (asl), and the local relief between the Washakie Range and western Owl Creek Mountains, Wind River Range, and Wind River Basin was 2.3 ± 0.8 km during the early Eocene. Post-Laramide regional uplift of ~1 km is required to form the present landscape in west-central Wyoming, which was most likely caused by regional Miocene–Pliocene uplift and exhumation.

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