



Middle Cenozoic diachronous shift to eolian deposition in the central Rocky Mountains: Timing, provenance, and significance for paleoclimate, tectonics, and paleogeography

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ABSTRACT

Eolian sedimentation was widespread in the Rocky Mountains (Rockies herein) during the middle and late Cenozoic. Although changes to eolian depositional environment have significance for tectonics, paleoclimate, and paleogeography, little is known regarding the timing of initiation and the provenance of these eolian sedimentary rocks in the Rockies. Here we study the timing of a transition to an eolian depositional environment in the central Rockies and the adjacent Great Plains during the middle Cenozoic, and use sandstone petrography and detrital zircon U-Pb geochronology to constrain the provenance of the eolian sedimentary rocks. Our samples have compositions of $Qm_{44}F_{27}Lt_{29}$ (Qm is monocrystalline quartz, F is feldspar, and Lt is total lithics), $Q_{47}F_{27}L_{26}$ (Q is total quartz, L is total non-quartzose lithics), and $Qm_{64}P_{28}K_8$ (P is plagioclase and K is potassic feldspar), and zircon age populations of 17–44 Ma, 45–218 Ma, 220–708 Ma, 948–1326 Ma, 1332–1816 Ma, and 1825–3314 Ma. The youngest zircon population was derived from distal volcanism in western and southwestern North America, and the other populations were derived directly from local Precambrian basements on Laramide ranges and recycled from Paleozoic–lower Cenozoic strata distributed along flanks of Laramide ranges and on the Sevier hinterland. The maximum depositional ages, based on the mean U-Pb ages of the youngest clusters of zircon grains, are generally consistent with the available ash radiometric dates for the latest Eocene–early Miocene samples, confirming that detrital zircon maximum depositional ages can be used to constrain depositional ages when ash beds or dateable minerals in ash beds were not present and when synchronous magmatic activity was intense. The occurrence of eolian deposition initiated during the latest Eocene–early Oligocene and became younger eastward, suggesting eastward progressive drying in the central Rockies. The diachronous drying may have resulted from the combined effect of renewed uplift of the Cordillera hinterland and central Rockies during the late Eocene and global cooling at the Eocene–Oligocene boundary. The provenance data presented here suggest that during the latest Eocene–early Oligocene, the westerlies and possibly the dry summer monsoon winds transported unlithified fluvial sediments and pyroclastic materials eastward and northeastward, and formed massive eolian deposits in the central Rockies and the adjacent

Great Plains. The eolian sedimentation continued into the Miocene and largely blanketed the Precambrian basement cores on Laramide ranges. The unlithified Oligocene eolian sediments were further eroded and recycled into the latest Oligocene–Miocene eolian sedimentary rocks.

INTRODUCTION

Ages, depositional environments, and provenances of eolian sedimentary rocks archive histories of sedimentation, dry climate, and atmospheric circulation, and thus have implications for tectonics, paleoclimate, and paleogeography. In the western interior of the U.S.A., eolian sedimentary rocks of middle and late Cenozoic age are widely distributed in the central Rocky Mountains (Rockies herein), southern Rockies, Great Plains, and Colorado Plateau as sand dunes or loess (e.g., Love, 1961; May and Russell, 1994; Tedford and Barghoorn, 1999; Honey and Izett, 1988; Hunt, 1990; Evanoff, 1993; Buffler, 2003; Cather et al., 2008), and may represent a permanent shift to the semiarid climate in the western interior.

On the Colorado Plateau and in the southern Rockies, eolian sedimentation initiated at the Eocene–Oligocene boundary (ca. 33.5 Ma) and ended ca. 27 Ma (May and Russell, 1994; Tedford and Barghoorn, 1999; Cather et al., 2008). Cather et al. (2008) proposed that the change to eolian deposition in the central Rockies should have been synchronous with that on the Colorado Plateau, and that both occurred at the Eocene–Oligocene boundary as a result of global cooling induced by glaciation in Antarctica. However, Evanoff et al. (1992) suggested that the transition appears younger eastward. Dating the initiation of the eolian deposition in the central Rockies will constrain the spatiotemporal pattern of the transition and help to resolve this debate. The eolian deposition in the central Rockies and the adjacent Great Plains lasted at least to the early Miocene and was referred as the White River and Arikaree aprons by Galloway et al. (2011). In addition to the ages, the provenance of the eolian sedimentary rocks in the central Rockies and the adjacent Great Plains has not been studied before. Previously researchers simply described these eolian sedimentary rocks as fine-grained volcanoclastic loess (Hunt, 1990; MacFadden and Hunt, 1998); this implies that the sediments were primarily derived from explosive

volcanism. Nevertheless, >85% of the zircon grains from the Oligocene eolian Chuska Sandstone in Arizona and New Mexico are of Proterozoic age, suggesting that the eolian sands were derived mainly from the Precambrian basement in central Arizona by southwesterly wind (Cather et al., 2008; Dickinson et al., 2010).

Here we study the provenance of Cenozoic eolian sedimentary rocks in the central Rockies and the adjacent Great Plains by applying detrital zircon U-Pb geochronology and sandstone petrography; and determine the age of the transition from fluvial- to eolian-dominated depositional environments by using maximum depositional ages based on the mean U-Pb ages of the youngest clusters of detrital zircon grains. The depositional ages of the oldest eolian deposits along a west-east transect are used to constrain whether the initiation of eolian deposition was synchronous or diachronous, and the provenance data are used to understand if the eolian sediments were primarily volcanoclastic or have been recycled. Our data also have significance for paleoclimate and paleogeography in the central Rockies and the adjacent Great Plains during the middle and late Cenozoic.

GEOLOGIC BACKGROUND

Tectonic Background

The central Rockies in Wyoming and northern Colorado are bounded by the Sevier fold-thrust belt to the west and the Great Plains to the east (Fig. 1). The modern topography of the central Rockies is characterized by intervening basement-cored mountains with crests that reach 4 km amsl (above mean sea level) and sedimentary basins with floors of 1.5–2 km amsl. Regional topography gradually decreases eastward to flat plains of ~1.0 km amsl in western Nebraska. The region was near sea level ca. 80 Ma, based on thick Cretaceous marine deposits (Roberts and Kirschbaum, 1995), and the subsequent Laramide orogeny formed the physiography of intervening sedimentary basins and mountain ranges during the latest Cretaceous–early Eocene (Dickinson and Snyder, 1978; Dickinson et al., 1988; DeCelles, 2004). By the end of the orogeny, the mountain ranges were as high as 3–4 km; however, basin floors remained near sea level (e.g., MacGinitie, 1969; Fan and Dettman, 2009; Fan et al., 2011).

The central Rockies and the adjacent Great Plains underwent differential uplift and eastward tilting during the middle and late Cenozoic. Currently, the timing and causes of the uplift are still debated. Stable and clumped isotope paleoaltimetry studies suggest that the regional relief between the central Rockies and the adjacent Great Plains was established before the Oligocene, and possibly resulted from uplift of the central Rockies with respect to the Great Plains during the late Eocene (Fan et al., 2014a, 2014b). This uplift could have been caused by lithospheric rebound induced by foundering of the Farallon flat slab or lower mantle lithosphere mantle beneath the central Rockies (Liu and Gurnis, 2010; Roberts et al., 2012), dynamic or isostatic

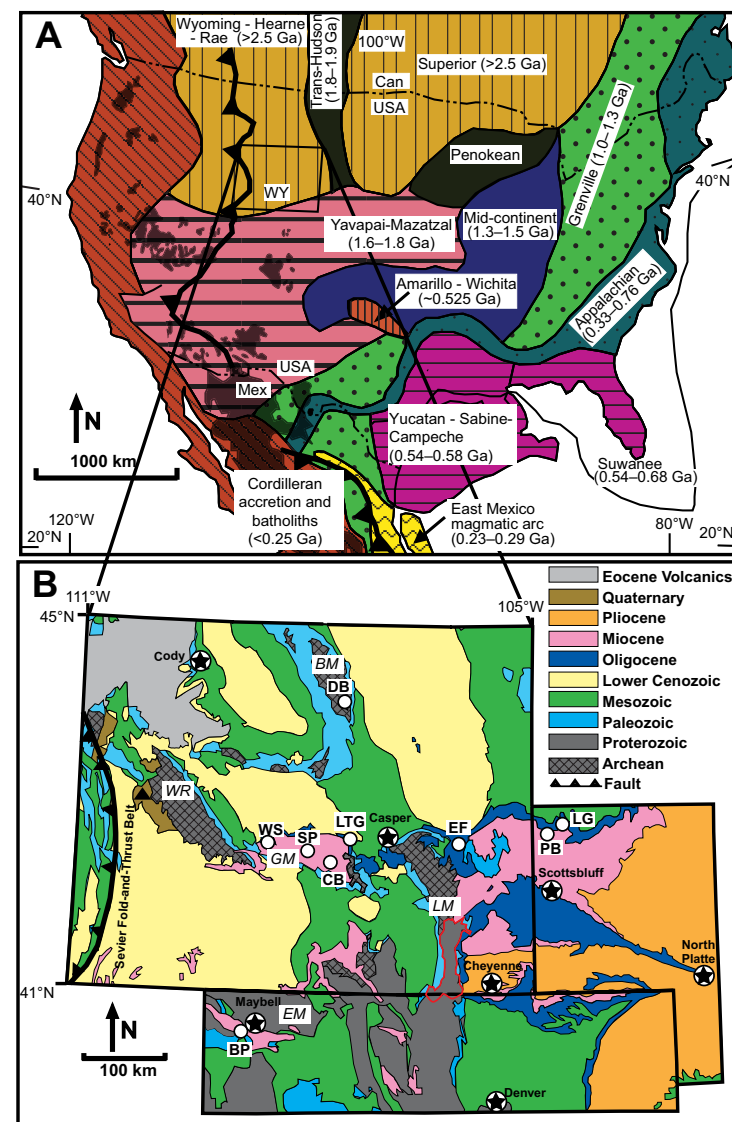


Figure 1. (A) Map showing locations of zircon provinces of various ages (after Dickinson and Gehrels, 2009). Locations of Cenozoic volcanic activity are represented by gray transparent polygons (after Ferrari et al., 2007; Lipman, 2007; Best et al., 2013). (B) Geological map of the study area in the central Rocky Mountains and adjacent Great Plains. White circles represent locations of the samples used for this study (see text). Red polygon highlights the locations of the Laramie anorthosite complex and Sherman batholith. Stars in circles represent major cities. LM—Laramie Mountains, WR—Wind River Range, BM—Bighorn Mountains, EM—Elkhead Mountains, GM—Granite Mountains.

topography induced by mantle and crustal anomalies (Lowry et al., 2000) and surface erosion, or a combination of these mechanisms. Other studies, based on erosional remnants of upper Neogene strata located on several Laramide mountain ranges, suggest that the intermontane basins in the central Rockies continually subsided and were buried to high levels until the latest Miocene (ca. 8 Ma) (Love, 1961; McKenna and Love, 1972; McMillan et al., 2006). This argument implies that the central Rockies underwent uplift during the latest Neogene, probably as a result of added upper mantle buoyancy (McMillan et al., 2006; Karlstrom et al., 2011). The latest Neogene uplift superimposed the localized extensional tectonics, which formed grabens on the Granite Mountains and the south end of the Wind River Range in central Wyoming (Love, 1970; Steidtmann and Middleton, 1991). The adjacent western Great Plains may have gained the present topography with the central Rockies during the late Eocene and the latest Neogene. However, Jones et al. (2015) suggested that the high topography of the Great Plains was established during the Laramide orogeny as a result of crustal dehydration.

Sedimentation and Erosion

Post-early Eocene strata are discontinuous and were deposited in fluvial and eolian depositional environments in the central Rockies (e.g., Love, 1961; McKenna and Love, 1972; Honey and Izett, 1988; Evanoff, 1993; Buffler, 2003). These strata have a spotty distribution on top of the Laramide mountain ranges as well as on floors of the intermontane basins (Fig. 1). The best documented erosional events in the region occurred during the late Eocene (ca. 43–37 Ma) and after the middle Miocene (Fig. 2). The late Eocene erosion event caused a regional depositional hiatus during the late Uintan–Duchesnean (Lillegraven, 1993), and incised deep canyons on several Laramide mountain ranges in the central Rockies (Steidtmann et al., 1989; Evanoff, 1990; Steidtmann and Middleton, 1991). The late Neogene erosion may have removed as much as 1.5 km of young sediment (McMillan et al., 2006; Pelletier, 2009). Although low-temperature thermochronology has the potential to elucidate post-Laramide erosion history, its applications to both mountain ranges and basin floors show cooling ages mostly during 80–40 Ma, coincident with the Laramide deformation, and a small magnitude of late Neogene erosional exhumation is only documented in a few study sites (see summary in Peyton and Carrapa, 2013).

These erosion events may result from climate change, tectonic uplift, or a combination of the two. The late Eocene erosion is most likely a result of differential tectonic uplift of the central Rockies with respect to the adjacent Great Plains (Cather et al., 2012; Fan et al., 2014a, 2014b), whereas the late Neogene erosion most likely resulted from climate change. Starting ca. 12 Ma, global temperature decreased gradually in response to Southern Hemisphere glaciation (Zachos et al., 2001). A significant amount of snowmelt runoff associated with the cooling excavated much of the upper Neogene strata in the central Rockies and deposited the late Miocene Ogallala Group in the Great Plains (Pelletier, 2009). However, tectonic uplift occurred ca. 8–6 Ma (McMillan et al., 2006; Duller et al., 2012), and may have caused additional erosion.

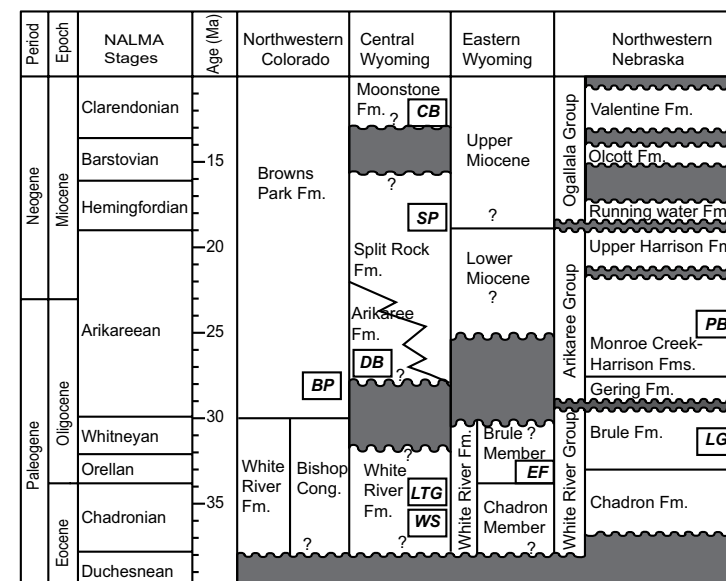


Figure 2. Chronostratigraphic framework of the studied latest Eocene–Miocene strata in Wyoming, northern Colorado, and western Nebraska. Stars represent the stratigraphic level of the detrital zircon samples. NALMA—North American land mammal age. Sample designations used for this study are in squares (see text).

SAMPLES

Middle and late Cenozoic sedimentary rocks distributed in Wyoming, northern Colorado, and western Nebraska generally comprised interbedded sandstone, conglomerate, loessite, and mudrock that were deposited in fluvial and eolian depositional environments. Compared to the fluvial sandstone, the eolian sandstones are massive, well sorted, and fine grained (e.g., Love, 1961; Hunt, 1990; Evanoff, 1993; Swinehart and Diffendal, 1987; LaGarry, 1998). Large-scale trough cross-stratifications were only documented in one location in the eolian sandstone in northwestern Colorado (Honey and Izett, 1988). We collected fine-grained sandstone and loessite samples from the latest Eocene–Oligocene White River Formation, late Oligocene–Miocene Arikaree and Browns Park Formations, and Miocene Split Rock Formation in the central Rockies, and the latest Eocene–Oligocene White River Group in western Nebraska (Figs. 1B and 2). The samples were collected from the type sections of these strata, or sections that were previously studied for ash geochronology, magnetostratigraphy, and mammal fossils, and have been assigned North American land mammal ages (NALMA). The location and available age constraints of each sample are summarized in Table 1.

TABLE 1. LOCATIONS AND AGE CONSTRAINTS OF THE SANDSTONE SAMPLES USED FOR DETRITAL ZIRCON GEOCHRONOLOGY IN THIS STUDY

Sample identification	Formation name	GPS location			NALMA stage	Age	
		Lat (°N)	Long (°W)	Elevation (m)		Radiometric age and magnetostratigraphy (Ma)	New detrital U-Pb* (Ma, 2σ)
WS 121.5	White River	42.76	108.27	1992	Uintan–Chadronian (Van Houten, 1964; Emry, 1975)	Chronos C16r–C16n (36.8–36.2) (Prothero and Sanchez, 2004)	36.0 ± 0.3 (n = 5)
LTG 560		42.64	106.76	1993	Chadronian (Emry, 1973)	⁴⁰ Ar/ ³⁹ Ar dating of Ash F: 35.72 ± 0.38 (biotite), 35.81 ± 0.09 (anorthosite) (Swisher and Prothero, 1990)	35.3 ± 1.0 (n = 5)
EF 15-60		42.61	105.25	1549	Orellan (Kron, 1978)	U-Pb dating of tuff 7: 32.9 ± 0.2 (zircon) (Scott and Bowring, 2000); ⁴⁰ Ar/ ³⁹ Ar dating of tuff 7: 30.7 ± 0.6 (biotite) (Swisher and Prothero, 1990)	33.0 ± 0.4 (n = 6)
LG 22-WB-8	Brule	42.85	103.58	1164	Whitneyan (Zanazzi et al., 2009)	⁴⁰ Ar/ ³⁹ Ar dating of Lower Whitney Ash: 31.2 ± 0.1 (biotite) (LaGarry, 1998)	31.6 ± 0.5 (n = 12)
PB 4583	Monroe Creek	42.74	103.81	1421	Early Arikareean (Hayes, 2005)	Chronos C9n–C7n (28.0–25.4) (MacFadden and Hunt, 1998; Hayes, 2005)	27.7 ± 0.8 (n = 5)
BP 98	Browns Park	40.61	108.33	1809	Arikareean and Hemingfordian (Honey and Izett, 1988)	K-Ar dating of a white ash: 24.8 ± 0.08 (biotite) (Izett et al., 1970)	27.5 ± 0.5 (n = 7)
DB 37	Split Rock	44.12	107.12	2730	Arikareean (McKenna and Love, 1972)	N.A.	26.6 ± 0.4 (n = 19)
SP 2		42.44	107.56	1939	Middle Hemingfordian (Munthe, 1988)	⁴⁰ Ar/ ³⁹ Ar dating of a ash 17.4 ± 0.08 (sanidine) (Izett and Obradovich, 2001); Chronos C5Br–C5Cr (17.4–15.8) (Luter et al., 2008)	18.8 ± 0.7 (n = 3)
CB 4	Moonstone	42.37	107.05	1924	Late Barstovian to early Hemphillian (Cassiliano, 2008)	U-Pb dating of lower tuff (LT-02): 11.3 ± 0.5 (zircon) (Scott, 2002)	27.3 ± 0.6 (n = 13)

Note: GPS—global positioning system; NALMA—North American land mammal age; Chron—magnetostratigraphic polarity chron. N.A.—not applicable.

*Based on results in this study; n—number of youngest zircon grains used for maximum depositional age.

SANDSTONE PETROGRAPHY

Methods

We studied 10 samples for sandstone petrography. Standard petrographic thin sections were stained for potassium feldspar and plagioclase. The thin sections were point counted by using a modified Gazzi-Dickinson method (Ingersoll et al., 1984), which considers grains larger than silt size (63 μm) within lithic fragments as individual grains (i.e., monocrystalline quartz). Because the grain sizes of most of the samples are fine, we reduce the grain size to be considered as individual grains to 50 μm. Modal framework grain compositions were determined by point counting 400 grains of each standard petrographic thin section. The point-counting parameters are listed in Table 2 and modal data are given in Table 3. Sandstone petrography data are displayed on a QFL (Q is total quartz, F is total feldspar, L is total non-quartzose lithics) ternary diagram for sandstone classification following Folk (1980) and on QmFLt (Qm is monocrystalline quartz and Lt is total lithics) and QmPK (P is plagioclase, K is potassium feldspar) ternary diagrams to assist in interpretation of sediment provenance following Dickinson and Suczek (1979) (Figs. 3 and 4).

Results

The studied sandstones are arkose, lithic arkose, and feldspathic litharenite (Figs. 3 and 4). The sandstones are all cemented by carbonate, and contain rounded quartz grains with occasional overgrowths and angular volcanic

TABLE 2. MODAL PETROGRAPHIC PARAMETERS

Symbol	Description
C	chert
Qm	monocrystalline quartz
Qp	polycrystalline quartz
Q	total quartz (Qm + Qp)
K	potassium feldspar
P	plagioclase
F	total feldspar (K + P)
Lsh	mudstone lithic
Lc	carbonate lithic
Lm	metamorphic lithics
Lvi	volcanic lithics (mafic and felsic)
Lvv	vitric volcanic lithics
Lv	total volcanic lithics (Lvi + Lvv)
Ls	total sedimentary lithics (Lsh + Lc + C)
Lt	total lithics (Ls + Lv + Lm + Qp)
L	total non-quartzose lithics (Ls + Lc + Lv)

glass and feldspar grains (Fig. 3). The intergranular volumes are large (Fig. 3), higher than 27%, indicating that the cement was formed during early diagenesis (Fan et al., 2014b). Monocrystalline quartz is the primary constituent of the framework grains. Other framework grains include polycrystalline quartz, chert, potassium feldspar, plagioclase, volcanic lithic fragments, and sedimentary lithic fragments (Fig. 3). Accessory framework grains include magnetite,

TABLE 3. MODAL PETROGRAPHIC DATA

Sample	Qm (%)	Qp (%)	C (%)	K (%)	P (%)	Lsh (%)	Lc (%)	Lvv (%)	Lvi (%)	Lm (%)	O* (%)
WS 208	38.5	0.0	5.0	6.0	26.3	0.8	0.5	11.8	7.8	0.0	3.5
LTG 740	45.0	1.0	4.0	9.0	13.0	0.8	0.0	10.5	13.0	0.0	3.8
EF 15-90	47.0	1.3	3.3	7.3	17.3	0.0	0.0	13.3	7.3	0.0	3.5
LG 22-20	50.8	1.3	3.5	10.0	28.5	2.3	0.8	0.0	1.3	0.0	1.8
LG 22-34	46.3	0.5	5.0	1.0	15.0	1.5	0.3	20.3	7.8	0.0	2.5
PB 4636	39.8	1.5	3.8	3.8	15.5	0.8	0.8	7.5	21.5	0.0	5.3
CB R4	48.5	0.5	4.0	7.8	25.3	1.5	0.3	7.0	2.8	0.3	2.3
BP 65	44.5	3.0	10.5	2.5	26.8	0.0	1.0	0.0	9.5	0.3	2.0
DB 16	35.8	2.5	4.3	3.3	12.5	1.8	0.5	22.8	12.3	0.0	4.5
SP 20	46.8	1.8	6.0	7.0	19.8	0.3	0.3	5.5	10.5	0.0	2.3

Note: See Table 2 for parameter abbreviations.

*O—other framework grains including biotite, hornblende, glauconite, magnetite, and zircon.

Figure 3. Photographs (under transmitted light) of petrographic thin sections of the representative eolian sandstone samples in the central Rocky Mountains. (A) DB 16 (latest Oligocene). (B) LTG 740 (latest Eocene). C—chert; P—plagioclase; Q_m—monocrystalline quartz; V—volcanic glass. Scale bars = 100 μ m.

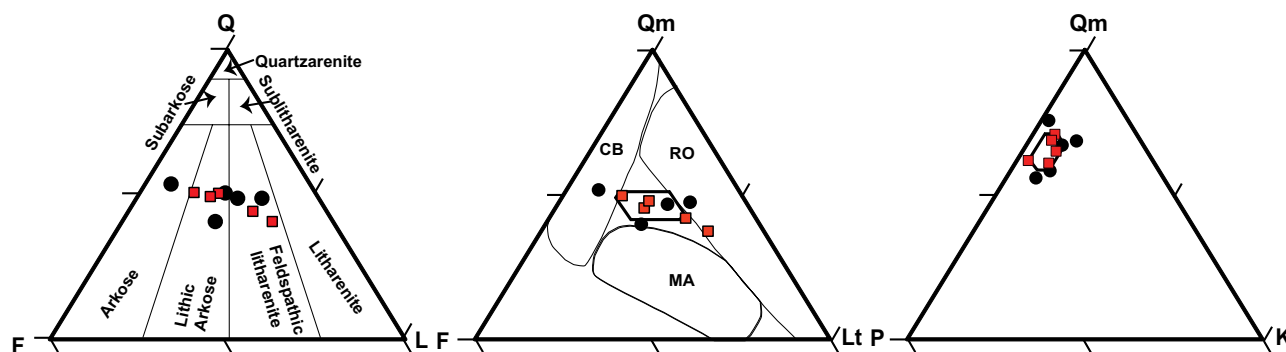
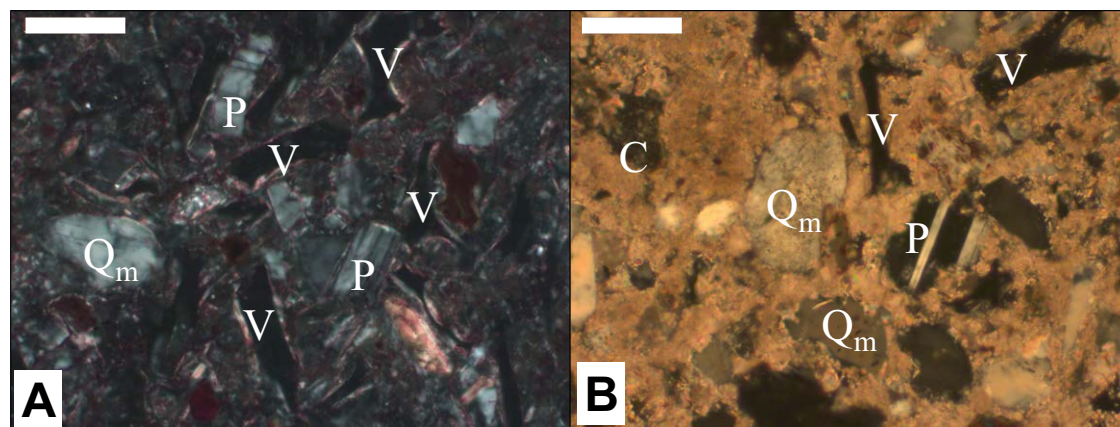


Figure 4. Ternary diagrams showing modal framework grain compositions of middle and late Cenozoic eolian sandstones. Lithology identification is after Folk (1980). Note that because the modified Gazzi-Dickinson method considers grains larger than silt size (63 μ m) within lithic fragments as individual grains, the data should contain less quartz and feldspar if the counting method in Folk (1980) is used. Provenance fields are after Dickinson and Suczek (1979). RO—recycled orogen; CB—continental block; MA—magmatic arc. Modal petrographic parameters and data are listed in Tables 2 and 3. Black circles are latest Eocene-Oligocene samples. Red squares are Miocene samples. Hexagons indicate one standard deviation uncertainties for average compositions of all samples.

biotite, zircon, hornblende, and glauconite. Volcanic lithic fragments dominate the lithic population and plagioclase dominates the feldspar population of the eolian sandstones (Fig. 4). The eolian sandstones have average modal compositions of $\text{Qm}_{44}\text{F}_{27}\text{Lt}_{29}$, $\text{Q}_{47}\text{F}_{27}\text{L}_{26}$, and $\text{Qm}_{64}\text{P}_{28}\text{K}_8$ (Fig. 4), and the compositions of the latest Eocene–Oligocene and Miocene eolian sandstones do not vary significantly (Fig. 4).

Interpretation

Sandstone modal composition reflects sandstone maturity and tectonic setting of deposition (Dickinson and Suczek, 1979; Folk, 1980; Dickinson et al., 1983). Because we modified the lower limit of grain size used for point counting to 50 μm , our data may reflect slightly more mature compositions compared to the data point counted using the Gazzi-Dickinson method. Our sandstone modal petrographic data show that the middle and late Cenozoic eolian sandstones were derived from mixed recycled orogen, continental block, and volcanic arc provenances (Fig. 4). Recycled orogen and continental block sources in the study area include the Precambrian basement rocks and the Phanerozoic strata distributed along flanks of the Laramide mountain ranges (Love and Christiansen, 1985). Phanerozoic strata distributed in the hinterland of the Sevier fold-thrust belt may also be a source of the eolian samples in western Wyoming and Colorado. Middle and late Cenozoic volcanic activities in western and southwestern North America (Best and Christiansen, 1991; McIntosh et al., 1992; Cunningham et al., 2007; Ferrari et al., 2007; Lipman, 2007; Best et al., 2013) supplied sediments that belong to magmatic arc provenance.

Rounded quartz grains with occasional overgrowths are indicative of reworked detrital grains from the underlying quartzose sedimentary units (Folk, 1980; Ingersoll et al., 1987). Polycrystalline quartz grains are derived from metamorphic basement and formed due to grain deformation under tectonic strain and low-grade metamorphism (Folk, 1980). These grains are most likely recycled from the Precambrian cores of Laramide ranges. Chert has possibly been recycled due to its chemical and mechanical durability, and it is therefore difficult to pinpoint to its original source (Folk, 1980). Plagioclase and potassium feldspars can be recycled from the local Archean cores of Laramide ranges (Love and Christiansen, 1985), or directly sourced and/or recycled from the Eocene–Miocene pyroclastic materials produced during volcanism in western and southwestern North America (Best and Christiansen, 1991; McIntosh et al., 1992; Cunningham et al., 2007; Ferrari et al., 2007; Lipman, 2007; Best et al., 2013). Cenozoic volcanism in western and southwestern North America is characterized by rhyolitic eruptions (Best and Christiansen, 1991; Best et al., 2013), which may bring abundant plagioclase and volcanic lithic grains into the sediments. These Cenozoic volcanic sources include, but are not limited to, the Absaroka and Challis volcanic fields, the Great Basin ignimbrite province, volcanic activity in Colorado, Idaho, Arizona, New Mexico, Utah, and the Sierra Madre Occidental in western Mexico (Moye et al., 1988; Best and Christiansen, 1991; McIntosh et al., 1992; Cunningham et al., 2007; Ferrari et al., 2007;

Lipman, 2007; Best et al., 2013). The presence of clastic sedimentary lithics suggests recycling of local Paleozoic, Mesozoic, and lower Cenozoic clastic sedimentary rocks. Overall, the latest Eocene–Miocene eolian sandstones in the central Rockies and adjacent Great Plains have low maturity, and the sources of the framework grains include proximal Laramide mountain ranges, Sevier hinterland, and distal volcanism in western and southwestern North America.

DETRITAL ZIRCON GEOCHRONOLOGY

Methods

A total of nine fine-grained sandstone and loessite samples were studied for detrital zircon U–Pb geochronology. Samples were collected and processed by standard methods for separating zircons. Four samples (WS121.5, LTG 171, EF 15-60, LG 6-29) were collected stratigraphically immediately above the regional transition from fluvial sedimentation to eolian sedimentation. U–Pb geochronology of zircons was conducted by a laser ablation–multicollector–inductively coupled plasma–mass spectrometry (LA-MC-ICP-MS) at the Arizona LaserChron Center (Gehrels et al., 2006, 2008). The analyses involve ablation of zircon with a Photon Machines Analyte G2 excimer laser, using a spot diameter of 35 μm . For samples with smaller grain sizes, the laser beam size was reduced to 25 μm in diameter. The ablation pit is ~15 μm in depth. Zircon grains were randomly selected for analysis to avoid bias to size or shape. The ablated material is carried in helium into the plasma source of a Nu Instruments Ltd high-resolution ICP-MS, which is equipped with a flight tube of sufficient width so that U, Th, and Pb isotopes are measured simultaneously. All measurements are made in static mode using Faraday detectors with $3 \times 10^{11} \Omega$ resistors for ^{238}U , ^{232}Th , ^{208}Pb , ^{207}Pb , and ^{206}Pb , and discrete dynode ion counters for ^{204}Pb and ^{202}Hg . Ion yields are ~0.8 mV per ppm. Each analysis consists of one 15 s integration on peaks with the laser off (for backgrounds), 15 s integrations with the laser firing, and one 30 s delay to purge the previous sample and prepare for the next analysis.

Common Pb correction was accomplished by measuring $^{206}\text{Pb}/^{204}\text{Pb}$ and assuming an initial Pb composition from Stacey and Kramers (1975) and uncertainties of 1.0 for $^{206}\text{Pb}/^{204}\text{Pb}$ and 0.3 for $^{207}\text{Pb}/^{204}\text{Pb}$. Subtraction of ^{204}Hg was accomplished by using the measured ^{202}Hg and natural $^{202}\text{Hg}/^{204}\text{Hg}$ ratio (4.34). Uncertainty in this value of $^{202}\text{Hg}/^{204}\text{Hg}$ is not significant because of the low intensities of Hg observed. Fractionation of $^{206}\text{Pb}/^{238}\text{U}$ and $^{206}\text{Pb}/^{207}\text{Pb}$ during ablation was monitored by analyzing fragments of a large concordant zircon crystal from Sri Lanka that has a known age of 563.5 ± 3.2 Ma (Gehrels et al., 2008). The uncertainty arising from this calibration correction, combined with the uncertainty from decay constants, age of the primary standard, and common Pb composition, contributes a 1% systematic error to the $^{206}\text{Pb}/^{238}\text{U}$ and $^{206}\text{Pb}/^{207}\text{Pb}$ ages (2σ). Zircon standard R33 was used as a secondary standard to ensure that data were reliable. For all samples reported herein, the average age of the R33 analyses was within 2% of the known age.

Supplemental Data table showing detrital zircon U-Pb geochronologic analyses. The table includes columns for sample ID, age (Ma), and discordance (%). The data is organized into rows for different samples, with some rows highlighted in green and others in yellow.

¹Supplemental Data. Detrital zircon U-Pb geochronologic analyses. Please visit <http://dx.doi.org/10.1130/GES01218.S1> or the full-text article on www.gsapubs.org to view the Supplemental Data.

The $^{206}\text{Pb}/^{207}\text{Pb}$ ages are used for grains older than 900 Ma, and $^{206}\text{Pb}/^{238}\text{U}$ ages are used for grains younger than 900 Ma. Grain ages were filtered by 30% discordance. Broader discordance tolerances to 50% were used for younger grains (younger than 100 Ma) due to the inherent inaccuracy in the measurement of $^{207}\text{Pb}^*$ in young grains, which often exceeds the analytical uncertainty of the measurement. Age groups were determined by identifying three or more grains with overlapping $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ages in the aggregate data set. After filtering the data for discordance, normalized relative age probability diagrams were constructed from concordant data (Fig. 5). Analyses that yielded isotopic data of acceptable discordance, in-run fractionation, and precision are shown in the **Supplemental Data**¹.

Results

Detrital zircon geochronology data show that zircon grains in the studied sediments have ages ranging between Archean (3314 Ma) and Miocene (17 Ma); 767 concordant ages are discussed here. Six zircon populations are defined based on grouping relative age peaks observed in the normalized age probability diagram (Fig. 5), including late Eocene–middle Miocene (44–17 Ma) zircons of population F, Late Triassic–middle Eocene (218–45 Ma) zircons of population E, Neoproterozoic–Middle Triassic (708–222 Ma) zircons of population D, Mesoproterozoic–Neoproterozoic (1326–948 Ma) zircons of population C, Paleoproterozoic–Mesoproterozoic (1816–1332 Ma) zircons of population B,

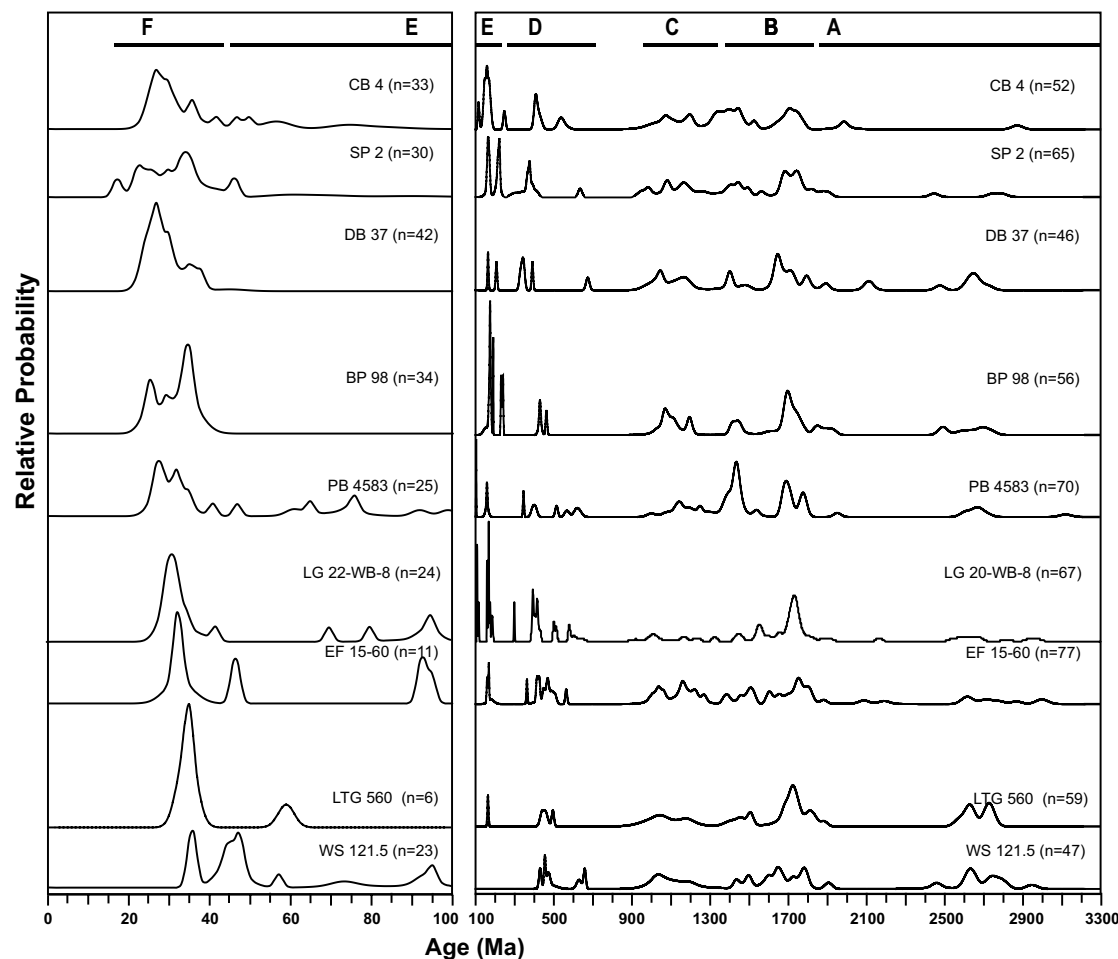


Figure 5. Normalized detrital zircon U-Pb age probability plots for the nine Cenozoic eolian sandstone samples.

and Archean–Paleoproterozoic (3314–1825 Ma) zircons of population A. Populations A and B are subdivided into subpopulations of A1, A2, A3, B1, B2, and B3, respectively (Table 4). The majority of the studied zircon grains are of populations B, C, and F (Fig. 6).

Interpretation

Major Detrital Zircon Sources

Major sources include populations B and C, derived predominantly from the Yavapai–Mazatzal and Grenville orogenic belts, and population F, derived from late Eocene–Miocene volcanic activity in western and southwestern North America (Fig. 1A). Population B includes three subpopulations (B1–B3). Zircons of B1 subpopulation are 1825–1590 m.y. old. Magmatic sources of this age are distributed currently in Arizona, New Mexico, Colorado, and northern Mexico and were formed during the Yavapai–Mazatzal orogeny (Fig. 1) (Condie, 1982; Grambling et al., 1988; Hoffman, 1988, 1989; Karlstrom et al., 2004; Nourse et al., 2005; Amato et al., 2008; Daniel et al., 2013; Jones et al., 2013). These basement rocks form cores of Laramide ranges south of the Archean–Proterozoic Wyoming shear suture zone (Condie, 1982), particularly in western Colorado and New Mexico (Jones and Thrane, 2012; Daniel et al., 2013; Jones et al., 2013). The Mojave basement rocks exposed in southern California and Arizona are of the same age and could also contribute zircons of subpopulation B1 (Miller et al., 1992; Barth et al., 2009). Subpopulation B2 (1589–1500 Ma) is a minor component and is originally from the Gawler craton in Australia (Fanning et al., 1988). Zircons of subpopulation B3 (1499–1332 Ma) were formed by anorogenic magmatism in south-central Laurentia (Anderson, 1983; Amato et al., 2008) and Sonora (Anderson and Silver, 1981; Nourse et al., 2005). Approximately 70% of Proterozoic magmatism occurred during 1499–1410 Ma, and the volume of magmatism decreased during 1410–1340 Ma (Anderson, 1983; Snoke, 1993). Midcontinental granitic intrusions of this age also underlie the Great Plains (Anderson, 1983; Bickford et al., 1986; Hoffman, 1988, 1989;

Karlstrom et al., 2004; Amato et al., 2008), and are in northern New Mexico (Daniel et al., 2013). Sporadic local exposures of magmatism of 1499–1332 m.y. old are limited to the Laramie anorthosite complex (1440 Ma) and Sherman batholith (1430 Ma) in southeastern Wyoming (Snoke, 1993). Subpopulations B1–B3 are all common in Mesoproterozoic–lower Cenozoic strata in the western U.S.A. (Gehrels et al., 1995; Gehrels and Stewart, 1998; Stewart et al., 2001; Dickinson and Gehrels, 2003, 2009; Fan et al., 2011; Dickinson et al., 2012; May et al., 2013). Therefore, zircons of population B were transported into the studied middle and upper Cenozoic strata by recycling Paleozoic–lower Cenozoic sedimentary rocks distributed along flanks of Laramide ranges and in the Sevier hinterland. Zircons of subpopulation B3 can also be transported into the studied strata in western Nebraska by direct weathering of the Laramie anorthosite complex and Sherman batholith.

Mesoproterozoic Grenvillian grains of 1330–948 Ma compose population C. This population was originally derived from the basement formed by the Grenville orogeny in a belt along the eastern flank of Laurentia (Fig. 1A; Dickinson and Gehrels, 2003). Although volumetrically less significant, the Pikes Peak batholith (1100–1000 Ma) of Colorado may also contribute zircons to this population (Anderson, 1983). Grenville-aged zircons are abundant in the Phanerozoic sedimentary rocks due to high zircon fertility (Moecher and Samson, 2006; Fan et al., 2011; Dickinson et al., 2012; May et al., 2013). Therefore, these zircon grains may have been mixed into the Cenozoic sedimentary rocks by eroding Paleozoic–lower Cenozoic sedimentary rocks.

Zircons of population F were formed by widespread Cenozoic volcanic activity in southwestern and western North America. Volcanism began in the northern portion of the Great Basin ca. 44 Ma and migrated southward and eastward across Nevada by Miocene time (Best and Christiansen, 1991). Profuse volcanism events in multiple caldera sources in western and southwestern North America were possible contributors to the population, including the Great Basin peak volcanism at 31–20 Ma (Best and Christiansen, 1991), the four ignimbrite pulses of the Mongollon–Datil field in New Mexico at 36.2–34.3 Ma, 32–31.4 Ma, 29.1–27.3 Ma, and 24.3 Ma (McIntosh et al., 1992), the

TABLE 4. PERCENTAGE OF ZIRCONS IN EACH POPULATION

Sample	A1 (3314–2491 Ma)	A2 (2490–1920 Ma)	A3 (1919–1825 Ma)	B1 (1824–1590 Ma)	B2 (1589–1500 Ma)	B3 (1499–1332 Ma)	C (1326–948 Ma)	D (708–222 Ma)	E (218–45 Ma)	F (44–17 Ma)	A1/ (A–F)	A1/ (A1–E)
WS 121.5	15.7	2.9	1.4	14.3	2.9	5.7	10.0	22.9	10.0	14.3	15.7	18.3
LTG 560	23.1	0.0	3.1	29.2	3.1	6.2	7.7	3.1	7.7	16.9	23.1	27.8
EF 15-60	10.2	3.4	1.1	21.6	3.4	6.8	6.8	9.1	13.6	23.9	10.2	13.4
LG 22-WB-8*	8.9	1.1	2.2	22.2	4.4	2.2	21.1	14.4	15.6	7.8	8.9	9.6
PB 4583	7.4	1.1	0.0	17.9	2.1	21.1	16.8	12.6	7.4	13.7	7.4	8.5
BP 98	5.6	3.3	3.3	16.7	1.1	4.4	30.0	15.6	6.7	13.3	5.6	6.4
DB 37	6.8	3.4	1.1	15.9	0.0	4.5	46.6	3.4	5.7	12.5	6.8	7.8
SP 2	2.1	1.1	3.2	16.8	1.1	9.5	26.3	12.6	11.6	15.8	2.1	2.5
CB 4	1.2	2.4	0.0	10.6	1.2	14.1	28.2	21.2	8.2	12.9	1.2	1.4

*Only one grain is not included in any of the populations.

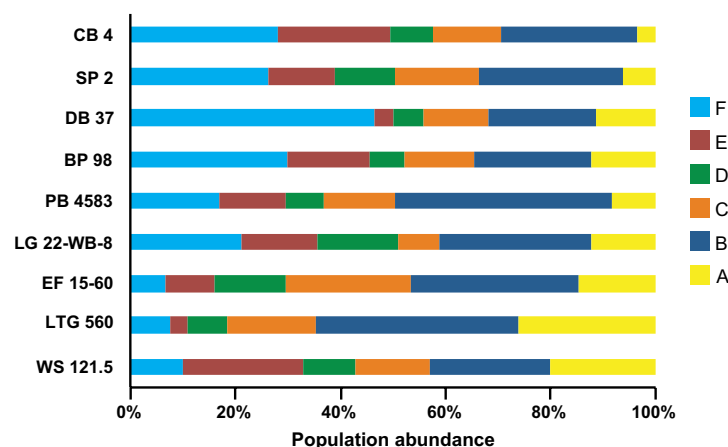


Figure 6. Relative abundance of defined zircon populations for the nine Cenozoic eolian sandstone samples. Data are summarized in Table 4.

two ignimbrite pulses of the Sierra Madre Occidental in Mexico at 32–28 Ma and 24–20 Ma (Ferrari et al., 2007), the peak volcanism of the southern Rocky Mountain field in Colorado at 29–27 Ma (Lipman, 2007), volcanism of the Marysville volcanic field in Utah at 32–17 Ma (Cunningham et al., 2007), and volcanism of the Snake River Plain and Yellowstone hotspot during the past ~17 m.y. (Tolan et al., 1989). Volcanism waned after ca. 20 Ma in the western U.S.A. (Best and Christiansen, 1991); this is supported by sparse zircon grains younger than 20 Ma in the two Miocene samples (SP 2 and CB 4). Zircon grains of population F were derived from direct ash fall out as well as erosion of unlithified sediments deposited broadly in western North America during the latest Eocene–Miocene.

Minor Detrital Zircon Populations

Population A (3310–1830 Ma) includes three subpopulations (A1–A3). Subpopulation A1 includes only zircons of Archean age (3310–2570 Ma). Archean terranes, including the Wyoming, Hearne, Rae, and Superior provinces, are exposed in western and central North America (Fig. 1A) (Whitmeyer and Karlstrom, 2007). The study area overlies the Archean Wyoming craton, outcrops of which make the cores of major Laramide ranges in Wyoming (Chamberlain et al., 2003). Zircons of subpopulation A2 (2490–1920 Ma) were formed by igneous activity in Canada (Hoffman, 1989); 2490–1920 Ma was a period of tectonic quiescence in the U.S.A. Zircons of subpopulation A3 (1919–1830 Ma) were formed by the Trans-Hudson orogeny, which represents the collision of the Wyoming Archean craton, the Superior province, and the Hearne-Rae province in Canada (Fig. 1) (Whitmeyer and Karlstrom, 2007). Zircons of all the

three subpopulations have been found in a wide range of Proterozoic–lower Cenozoic strata in North America (e.g., Stewart et al., 2001; Nourse et al., 2005; Gleason et al., 2007; Gehrels et al., 2011; Fan et al., 2011; May et al., 2013). Zircons of subpopulation A1 could be mixed into the middle and late Cenozoic strata by direct erosion of Laramide basement-cored ranges in Wyoming or recycling from Proterozoic–lower Cenozoic sedimentary rocks. However, A2 and A3 can only be recycled from older sedimentary rocks. Population A is dominated by subpopulation A1 with minor contributions of subpopulations A2 and A3, suggesting that direct erosion of Laramide basement-cored ranges provided most of the zircons of population A into the studied strata.

Population D comprises zircon grains 708–222 m.y. old, which were mainly formed by the Appalachian orogeny (640–265 Ma) and a Permian–Triassic arc in eastern Mexico (284–232 Ma) (Torres et al., 1999; Eriksson et al., 2003). Tectonomagmatic events of the Appalachian orogeny include the Avalonian–Carolinian at 640–580 Ma, Potomac ca. 500 Ma, Taconian at 500–430 Ma, Acadian at 400–350 Ma, and Alleghanian at 325–265 Ma (Eriksson et al., 2003). Other sources of grains of these ages include the Suwannee terrane in the subsurface of Florida, the Yucatan block in Mexico, and the Cambrian Wichita igneous province in southern Oklahoma (Mueller et al., 2013; Hanson et al., 2013). An additional source is the magmatism associated with the Antler–Sonoma orogeny (450–330 Ma) in Nevada and Idaho (Dickinson, 2004). However, none of the sources are exposed in the vicinity of the study area. This population is found in Paleozoic–lower Cenozoic strata in the western and central U.S.A. (Gehrels et al., 1995; Gehrels and Stewart, 1998; Stewart et al., 2001; Dickinson and Gehrels, 2003, 2009; Fan et al., 2011; Dickinson et al., 2012; May et al., 2013). Therefore, zircons of population D in the studied strata were probably recycled from Paleozoic–lower Cenozoic strata.

Population E comprises zircons 218–45 m.y. old and derived from the Cordilleran magmatic arc in the broad area of western North America (Chen and Moore, 1982; Barth and Wooden, 2006). Major episodes of Sierra Nevada magmatism occurred during the Jurassic (200–150 Ma) and Late Cretaceous (120–85 Ma) (Armstrong and Ward, 1993). Two major episodes of Baja California Peninsula magmatism occurred during the Early Cretaceous (135–115 Ma) and Late Cretaceous (100–80 Ma) (Armstrong and Ward, 1993). Southern Californian peak magmatism occurred during the Late Cretaceous (85–70 Ma) (Jacobson et al., 2011). Magmatism swept northeastward from the Sierra Nevada into central Montana from 80 to 65 Ma (Armstrong and Ward, 1993). Magmatism was generally sporadic in Wyoming and the nearby area ca. 70–45 Ma due to the low-angle subduction of the Farallon oceanic plate (Dickinson and Snyder, 1978). The sparse volcanism events that may have contributed zircons to studied samples includes the Colorado mineral belt (75–43 Ma), Black Hills (60–47 Ma), Kamlloops–Challis–Absaroka (55–40 Ma), and Rattlesnake Hills (ca. 44 Ma) (Armstrong and Ward, 1993). Zircons of this population are found in Triassic–lower Cenozoic strata in the western and central U.S.A. (Dickinson and Gehrels, 2003, 2009; Fan et al., 2011; May et al., 2013). Therefore, zircons of this population were mixed into the studied samples by recycling Triassic–lower Cenozoic strata.

In summary, zircon populations in the latest Eocene–Miocene eolian sediments in the central Rockies and adjacent Great Plains suggest that sources of these sediments include local Precambrian–lower Cenozoic rocks in the Laramide province and Sevier hinterland, and direct distal input of middle-late Cenozoic volcanism in the broad area of western and southwestern North America. Distal and local provenances represented by the zircon populations are consistent with the interpretation of the sandstone petrography results. The high abundance of recycled zircons suggests that these sediments are not entirely volcanoclastic as previously thought (Hunt, 1990; MacFadden and Hunt, 1998).

■ MAXIMUM DEPOSITIONAL AGES AND TIMING OF FLUVIAL-EOLIAN TRANSITION

Maximum depositional ages of all the samples were calculated from the weighted mean of more than three youngest grains using Isoplot (Ludwig, 2008), and reported with a mean square of weighted deviates (MSWD) (Fig. 7; Table 1). Although maximum depositional ages are of lower precision compared to depositional ages based on radiometric dating of ashes, they provide

good estimates of depositional ages when air-fall volcanic ashes or dateable minerals are not present, and when there are abundant syndepositional volcanic zircons (Fan et al., 2015). The maximum depositional ages of most of the studied samples are consistent with available radiometric, biostratigraphic, and magnetostratigraphic ages (Table 1). The consistency is expected because concurrent volcanic eruptions produced large volumes of ash, which contributed to the eolian sediments either as direct air-fall grains, or as recycled grains from slightly older, unlithified deposits covering western and southwestern North America during the middle and late Cenozoic. This finding confirms a previous study that suggests maximum depositional ages can be used to constrain depositional ages of strata when synchronous volcanism was intense, and ash beds and dateable minerals are not present (Fan et al., 2015). The only exception is the youngest sample, CB 4, which has a maximum depositional age ~16 m.y. older than the radiometric dates of two ash beds stratigraphically nearby (Scott, 2002). This is particularly intriguing because the slightly older sample, SP 2, contains syndepositional zircons ~17 m.y. old. The lack of late Miocene syndepositional zircons in sample CB 4 may be caused by the slowing of volcanic activity after 20 Ma in the western U.S.A. (Best and Christiansen, 1991), or because young ashes in western U.S.A. were trapped locally.

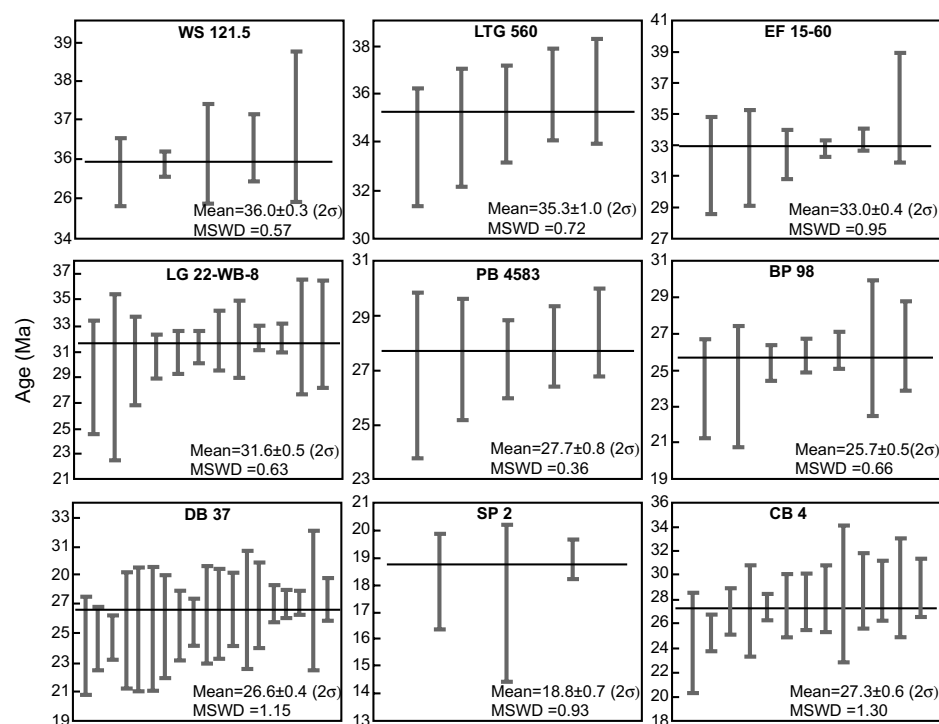


Figure 7. Maximum depositional ages of the latest Eocene–Oligocene sandstone samples. Error bars represent 2σ (MSWD — mean square of weighted deviates).

The four samples (WS121.5, LTG 560, EF 15-60, LG 22-WB-8) that represent the regional transition from fluvial sedimentation to eolian sedimentation from western Wyoming to western Nebraska have maximum depositional ages of 36.0 ± 0.3 Ma, 35.3 ± 1.0 Ma, 33.0 ± 0.4 Ma, and 31.6 ± 0.5 Ma (Fig. 7), showing an eastward-younging trend. Such a trend is also reflected by the compiled NALMA, radiometric, and magnetostratigraphic ages (Table 1), suggesting that the eolian deposition initiated during the latest Eocene–early Oligocene in the central Rockies and the adjacent Great Plains. The timing of the eastward initiation of eolian deposition documented in this study is generally consistent with the initiation of eolian sedimentation at the Eocene–Oligocene boundary in the southern Rockies and on the Colorado Plateau (Cather et al., 2008), suggesting that eolian deposition broadly covered the western interior during the Oligocene. This diachronous change of climate in the central Rockies is also recorded in the change of nonmarine gastropod fauna that reflects west to east progressive drying and a change from a moist subtropical to a semiarid climate (Evanoff et al., 1992).

COMPARISON OF PROVENANCE DATA

The nine eolian samples studied for detrital zircon geochronology can be divided into two groups based on their depositional ages (Fig. 8). The latest Eocene–early Oligocene (E–O) group includes WS 121.5, LTG 560, EF 15-60, and LG 33-WB-8, and the latest Oligocene–Miocene (O–M) group includes PB 4583, BP 98, DB 37, SP 2, and CB 4 (Table 1). Kolmogorov–Smirnov (K–S) tests were conducted to compare the provenance of the two groups of samples. P-values ≤ 0.05 indicate the grain populations of two compared samples from different sources with $\geq 95\%$ confidence, thus the provenances of the two samples are statistically distinguishable. K–S test results of the two groups show that the P value of the populations F–D (younger than 948 Ma) is <0.05 , but the P value of the populations C–A (older than 948 Ma) is 0.97, suggesting that zircon populations older than 948 Ma are of similar proportional contributions to the two groups of samples, but zircon populations younger than 948 Ma are of different proportional contributions. The major difference of populations D and E between the two groups of samples is that the samples in the O–M group have proportionally more early Mesozoic (240–210 Ma) grains than the samples in the E–O group (Figs. 6 and 8), which may suggest that lower Mesozoic siliciclastic sedimentary rocks became a more important source during the latest Oligocene–Miocene. The major difference of population F between the two groups of samples is that the samples in the O–M group are younger, and have abundant young grains in the range of 32–24 Ma. This age group matches the time of peak volcanism in the Great Basin, Sierra Madre Occidental, Mongol–Datil, and Marysvale fields and other small fields in the southern Rockies (Best and Christiansen, 1991; McIntosh et al., 1992; Cunningham et al., 2007; Ferrari et al., 2007; Lipman, 2007). This age group also occurs in the middle and late Cenozoic fluvial sedimentary rocks in the study area and represents high ignimbrite flux in western and southwestern North America (Fan et al.,

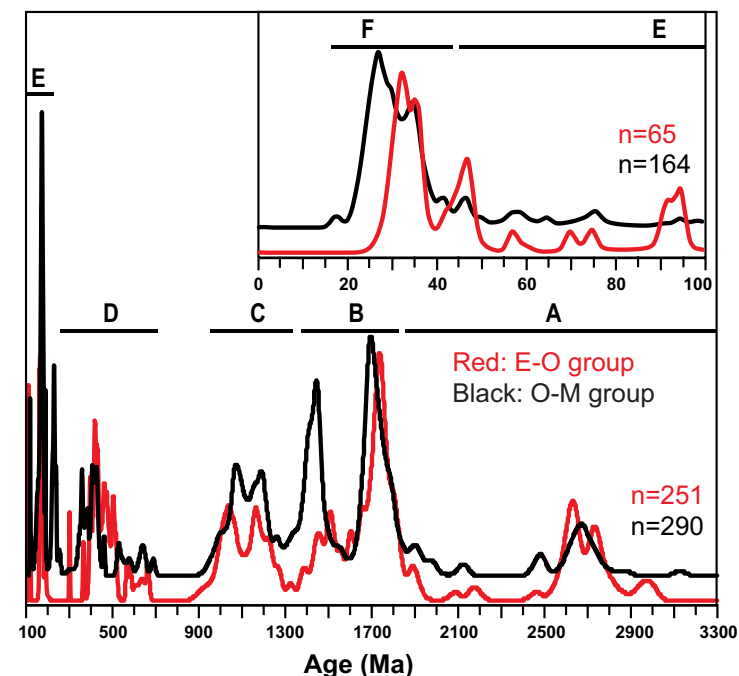


Figure 8. Comparison of normalized U–Pb age probability plots for the latest Eocene–early Oligocene (E–O) group and latest Oligocene–Miocene (O–M) group. The E–O group includes WS 121.5, LTG 560, EF 15-60, and LG 33-WB-8, and the O–M group includes PB 4583, BP 98, DB 37, SP 2, and CB 4.

2015). High abundance of population F during the latest Oligocene–Miocene may suggest that large volumes of unlithified late Eocene–Oligocene sediments were available for wind erosion and transport.

Samples from the CB, SP, and DB sites in the O–M group and samples from the LTG, WS, and EF sites in the E–O group are very close to or directly from strata covering the Laramide basement-cored ranges (Fig. 1), but the two groups of samples have different abundances of Archean zircons (Figs. 6 and 8; Table 4). The difference may result from partial burial of Archean basement during the latest Oligocene–Miocene. The other possibility is that the high abundance of population F in the O–M group diluted the Archean population. However, the abundance of subpopulation A1 (Archean) relative to populations A–E, not including population F, is 13%–28% for the three samples in the E–O group and 1%–8% for the three samples in the O–M group (Table 4). This difference may suggest that the high abundance of population F is not the reason for lower abundance of Archean zircons in the O–M group. We suggest that the burial of Archean basement on Laramide mountain ranges during the late Oligocene and Miocene was the result of widespread eolian deposition. The gradual burial

of Archean basement may also suggest that central Rockies did not undergo any uplift or exhumation during the latest Oligocene–early Miocene.

Although the number of zircons used to infer widespread eolian deposition during the latest Oligocene and Miocene is small, the interpretation is further supported by preservation of eolian sedimentary rocks on high mountain ranges and broad geographic areas (Fig. 9). On the basis of this study, eolian sedimentary rocks of the latest Eocene–early Oligocene age are distributed in central Wyoming and northwestern Nebraska. It may also extend to southeastern Wyoming, western Nebraska, and South Dakota based on distribution of the Brown Siltstone Beds of White River Group in Nebraska (Swinehart et al., 1985) and Poleside Member of the Brule Formation (Benton et al., 2015). Latest Oligocene–Miocene eolian sedimentary rocks, however, are also distributed in the Browns Park Formation in northwestern Colorado and south and central Wyoming (Honey and Izett, 1988; Buffler, 2003), and throughout western Nebraska as part of the Arikaree Group (Swinehart and Diffendal, 1987). The distribution may extend as far as North Dakota, where well-sorted volcanoclastic siltstones in the Arikaree Formation are documented (Murphy et al., 1993). In addition, Miocene eolian sedimentary rocks are distributed on mountains ~3 km above sea level, including Elkhead Mountain in northwestern Colorado and the Bighorn Mountains in northern Wyoming (McKenna and Love, 1972; Buffler, 2003).

SIGNIFICANCE FOR PALEOCLIMATE, TECTONICS, AND PALEOGEOGRAPHY

Paleoclimate and Implication for Tectonics

Eolian deposition is inherently tied to aridification and often initiates as a result of global cooling, local uplift, or a combination of the two (e.g., An et al., 2001; Liu et al., 2009; Sun et al., 2010). Cather et al. (2008) suggested that the

initiation of eolian deposition on the Colorado Plateau and the southern Rockies was caused by the global oxygen isotope excursion Oi-1 cooling event due to glaciation in Antarctica (Zachos et al., 2001). The maximum depositional ages of the four samples that represent the initiation of drying, as well as the compiled NALMA and radiometric ages, show that the transition to dry climate in the central Rockies and the adjacent Great Plains is diachronous rather than regionally synchronous. Global cooling as the sole cause for eolian deposition and drying would likely result in a regionally synchronous transition, and thus is not well supported by the data. Another possible mechanism of regional drying is the rain shadow effect caused by the renewed uplift of the central Rockies as well as the broad uplift in the Cordillera hinterland. Both the central Rockies and the Cordillera hinterland underwent at least 1 km of uplift during the late Eocene (Mix et al., 2011; Cather et al., 2012; Chamberlain et al., 2012; Fan et al., 2014a, 2014b). However, the high topography in the Cordillera hinterland and the central Rockies was formed by the late Eocene, thus the mechanism is difficult to invoke to explain the drying in western Nebraska during the early Oligocene. We suggest here that the combined effect of surface uplift and global cooling may have caused the eastward drying. The highlands in western North America may have blocked the moisture from the Pacific Ocean and caused leeward drying in western and central Wyoming during the late Eocene. Global cooling at the Eocene–Oligocene boundary superimposed on the eastward-decreasing topography in the study area reduced the moisture from the Gulf of Mexico to western Nebraska. This speculation of the combined effect of topography and global cooling can only be fully assessed when the spatiotemporal patterns of drying are compared with results of paleoclimate model simulations.

The change to eolian depositional environment did not occur synchronously across the region at the Eocene–Oligocene boundary, ca. 33.5 Ma, possibly because uplift of the Rockies and Cordillera hinterland suppressed climate

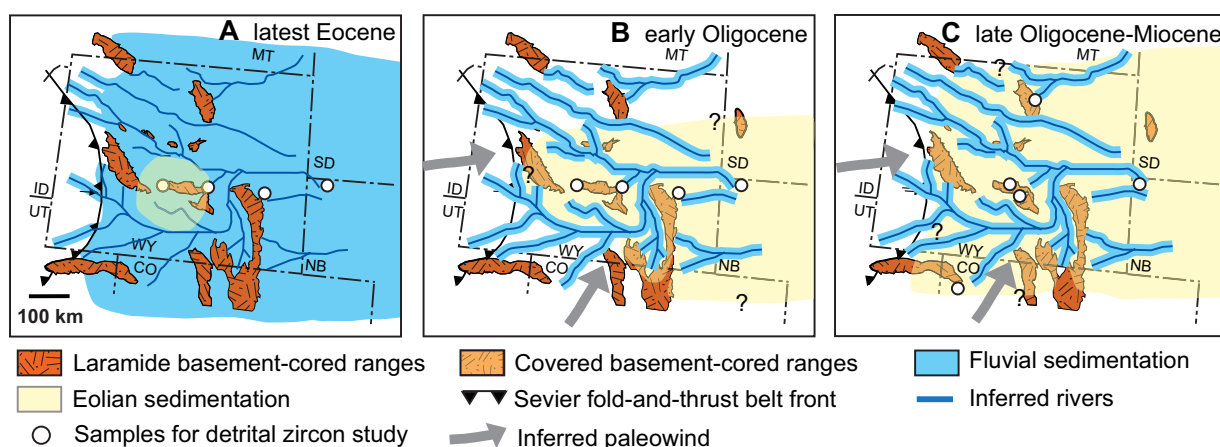


Figure 9. Paleogeographic interpretation of the central Rocky Mountains and adjacent Great Plains during the late Eocene–Miocene. (A) Initiation of eolian deposition in western Wyoming and widespread fluvial deposition in other area during the latest Eocene. (B) Coexisting eolian and fluvial deposition during the early Oligocene. Note that basement cores of Laramide ranges were largely exposed in A and B. (C) Widespread eolian deposition buried Laramide ranges during the latest Oligocene–early Miocene. The latest Eocene–Oligocene fluvial drainage pattern is based on Clark (1975), Seeland (1985), Lillegraven and Ostresh (1988), and Fan et al. (2015), and the early Miocene fluvial drainage pattern is assumed to be the same as the Oligocene.

and environmental changes in response to global cooling. In South Dakota and Oregon, where the uplift of Rockies has limited influence on atmospheric circulations, the ecosystem changed from tropical forest to grassy woodland and the paleoprecipitation amount declined significantly in response to global cooling (Retallack et al., 2004; Retallack, 2007; Sheldon, 2009). However, in the central Rockies and its vicinity, stable and clumped isotope studies do not document any major cooling or drying in Wyoming and western Nebraska across the boundary (Fan et al., 2014a, 2014b), the reconstructed paleoprecipitation amount declined slightly or remained low in Nebraska and Montana (Retallack et al., 2004; Retallack, 2007; Sheldon, 2009), and the ecosystem changed from grassland to savannahs and woodland in northeastern Colorado (Hembree and Hasiotis, 2007). These observations can be explained as local responses to the rain shadow effect of the Rockies high topography, which overprints global climate changes.

Paleogeography

Three factors, including a source for sediment generation, wind regime and energy for sediment transport, and location for sediment accumulation, must be met in order to form eolian sedimentation (Tsoar and Pye, 1987). Direct wind deflation of Precambrian basement and Paleozoic–lower Cenozoic strata potentially produce unlithified sediments. A typical example for direct wind deflation of bedrock is that the dry and strong Asian Winter Monsoon from Siberia caused significant erosion in northwestern China, which resulted in the Yardang landscapes and the deposition of loess on the Loess Plateau (e.g., Kapp et al., 2011). However, this mechanism requires strong wind energy and often forms eolian erosion remnant landscapes, which became extensive in Asia, North America, South America, Europe, and New Zealand during the past 2 m.y., and is tied to glacial periods when global climate became cold and dry with abundant loose glacial till, fluvial sediments, or desert distributed upwind for erosion (e.g., An et al., 1990; Muhs et al., 2003, 2008; Aleinikoff et al., 2008). No desert or glaciation existed in North America during the latest Eocene–early Miocene, thus glacial tills are not the potential sources of the eolian sedimentary rocks in this study.

We suggest that the sources of the eolian sedimentary rocks include distal middle and late Cenozoic pyroclastic materials and local late Eocene–early Miocene fluvial sedimentary fills in the Laramide intermontane basins. Fluvial systems flowed southward and eastward during the Oligocene in the central Rockies and the adjacent Great Plains (e.g., Clark, 1975; Seeland, 1985; Lillegraven and Ostresh, 1988). Such rivers cut large, deep valleys on the Wind River Range and Laramie Range during the late Eocene (Steidtmann et al., 1989; Evanoff, 1990; Steidtmann and Middleton, 1991). Fluvial deposits of the rivers were not lithified during the eolian deposition and could be easily deflated and transported by wind. An example of this is loose fluvial sediments deposited on the banks of the Colorado River in the southwestern U.S.A., providing a major source for wind transport and eolian sedimentation along the

river (Muhs et al., 2003). This interpretation is also supported by the similarity of detrital zircon age distribution between middle Cenozoic eolian and fluvial sedimentary rocks in the study area. By comparing the zircon age distributions of the eolian sandstones and the interbedded Oligocene and Miocene fluvial sedimentary rocks in the study area, Fan et al. (2015) suggested that the co-existing eolian and fluvial depositional processes in the central Rockies supplied loose sediments for each other and promoted sediment mixing during the late Paleogene. Moreover, the older eolian deposits may become an important source for eolian deposition during the latest Oligocene–Miocene.

Transport of the young zircons from the broad area of western and southwestern U.S.A. to the central Rockies does not require eastward- and northeastward-flowing transcontinental rivers, although the Sierra Madre Occidental volcanic field in Mexico is the largest volcanic field and may be the most important source of zircons in population F (Fan et al., 2015). In fact, there is no geologic evidence supporting the existence of such rivers. Transport of unlithified fluvial deposits in the central Rockies and pyroclastic sediments mantling the broad area of the western U.S.A. also does not require a strong wind regime. In Asia, the Siberian High is strengthened during wintertime and glacial periods, causing a strong Asian Winter Monsoon (e.g., Gong et al., 2001). Southeastward transport direction of eolian dusts in Asia is indicative of wind regime (An et al., 2001). In the central Rockies, because the studied period is earlier than the establishment of Northern Hemisphere glaciation during the Pliocene (Zachos et al., 2001), dry polar airflow might not be established during this studied period and cannot be the wind regime for the documented eolian sedimentation. The only documented wind flow direction in the central Rockies, based on large-scale trough cross-stratification in northwest Colorado, is northeastward (Honey and Izett, 1988; Buffler, 2003). Although pyroclastic sediments broadly covered the western and southwestern U.S.A. during the latest Eocene–Miocene, we think the component of young zircons directly ejected from volcanic eruptions to the central Rockies should be small, given the great transport distances. We speculate that most of the young zircons were transported eastward and northeastward by wind through several intermediate steps and transient storages. We suggest that dry upper troposphere westerlies may be the main wind flow that transported different grain sizes of pyroclastic from the broad Cordilleran hinterland and local fluvial deposits to form the studied eolian sediments in the central Rockies and the adjacent Great Plains. It is also possible that near-ground southwestern summer monsoons brought pyroclastic materials from southwestern North America to the studied area.

The interpretations of sediment provenance and inferred transport mechanism and wind regime are used to reconstruct the paleogeography of the study area (Fig. 9). During the latest Eocene–early Oligocene, widespread volcanism in western and southwestern North America and large rivers draining local Laramide highlands produced abundant loose materials for wind transport. Dry westerlies and near-ground monsoonal wind reworked these unlithified fluvial and pyroclastic materials and formed massive eolian sheets in the central Rockies and the western Great Plains. These eolian deposits

mantled Laramide mountain ranges gradually, and the Archean cores of Laramide ranges were largely buried during the latest Oligocene and Miocene, evidenced by Archean zircons being less abundant in the O-M group than in the E-O group. Although large fluvial valleys of late Oligocene–early Miocene age were not documented in the region, the basal Arikaree conglomerates were deposited in fluvial environments and are widely distributed in Wyoming and western Nebraska (e.g., McKenna and Love, 1972; Swinehart et al., 1985; Evanoff, 1993). These fluvial systems may erode and transport the late Eocene–early Oligocene eolian sediments and provided source materials for eolian sedimentation during the Miocene. This is further supported by the poor preservation of Oligocene eolian sediments in the central Rockies and high abundance of Oligocene zircon grains in the O-M group.

CONCLUSIONS

1. Eolian sedimentation started in the central Rockies and adjacent Great Plains during the middle Cenozoic. The eolian sediments have average modal compositions of $\text{Qm}_{44}\text{F}_{27}\text{Lt}_{29}$, $\text{Q}_{47}\text{F}_{27}\text{L}_{26}$, and $\text{Qm}_{64}\text{P}_{28}\text{K}_8$, with the framework grains derived from mixed recycled orogen, continental block, and volcanic arc sources. A total of 766 detrital zircon ages from 9 eolian sandstone samples show populations of 17–44 Ma, 45–218 Ma, 220–708 Ma, 948–1326 Ma, 1332–1816 Ma, and 1825–3314 Ma. The volcanic arc source is distal volcanism, which occurred broadly in western and southwestern North America during the middle and late Cenozoic. This source provided the youngest zircon population (44–17 Ma). The recycled orogen and continental block sources are Precambrian basement distributed on local Laramide ranges and Paleozoic–lower Cenozoic strata distributed on flanks of the ranges. These sources provided zircon populations older than 45 Ma. Phanerozoic strata distributed in the Sevier hinterland may also be a source of eolian deposition in western Wyoming and western Colorado.

2. Maximum depositional ages based on clusters of youngest detrital zircon U-Pb ages of four oldest middle Cenozoic eolian samples show that the onset of eolian deposition had an eastward-younging trend, beginning in central Wyoming ca. 36 Ma and western Nebraska ca. 30 Ma. The diachronous transition of depositional environment suggests eastward progressive drying, which may be caused by the combined effect of renewed uplift of the central Rockies and Cordillera hinterland during the late Eocene and global cooling at the Eocene–Oligocene boundary.

3. The maximum depositional ages of the latest Eocene–early Miocene samples are consistent with the available radiometric, magnetostratigraphic, and biostratigraphic ages; this confirms a previous study suggesting that maximum depositional ages can be used to constrain depositional ages of strata when synchronous volcanism was intense and ash beds and dateable minerals are not present.

4. Collectively, our presented data were used to reconstruct the paleogeography of the study area during the middle and upper Cenozoic. During the

late Eocene, Laramide ranges were eroded by fluvial processes and produced large amount of un lithified fluvial sediments. These loose fluvial sediments and pyroclastic materials derived from concurrent volcanism in western and southwestern North America were transported by dry westerlies toward the east and possibly by summer monsoon winds toward northeast. Eolian deposition broadly covered the western interior of the U.S.A. during the Oligocene and Miocene, and may have intensified and largely blanketed Laramide ranges in the central Rockies during the latest Oligocene–Miocene.

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