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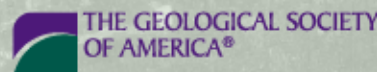
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Cenozoic collapse of the eastern Uinta Mountains and drainage evolution of the Uinta Mountains region

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ABSTRACT

Coupled detrital sanidine and zircon data, combined with sedimentological and stratigraphic observations, provide temporal constraints on the post-Laramide paleogeographic and structural evolution of the eastern Uinta Mountains region from the late Eocene to late Miocene (ca. 36–8 Ma). Maximum depositional ages (MDAs) calculated from detrital zircon U-Pb and detrital sanidine ⁴⁰Ar/³⁹Ar ages indicate that the most significant Paleogene fluvial system in the region, represented by the Bishop Conglomerate, existed from 36 to 27 Ma. The abundance of red sandstone and gray limestone clasts, paleocurrent directions, and the large number of Grenville-age detrital zircons suggest that the Uinta Mountain Group (UMG) facies of the Bishop Conglomerate are tributaries that flowed radially away from the crest of the Uinta Mountains. To the north of the Uinta Mountains, these rivers joined a mainstem river in southwestern Wyoming represented by the Bishop Conglomerate Firehole Canyon (FC) facies. This facies consists of rounded cobble-to pebble-sized quartzite clasts with minor quantities of volcanic rocks, has westward paleocurrent directions, and abundant young (younger than 40 Ma) detrital zircon and sanidine grains. Detrital sanidine age and geochemical data suggest that these young detrital grains are tephra that originated from the Basin and Range volcanic field, which was subsequently reworked into Bishop Conglomerate sediments. The more regional headwaters of the mainstem river could have been located east of the Uinta Mountains, or in the Challis and Absaroka volcanic fields and the Wind River Mountains located to the northwest of the region. The question of whether the FC facies of the Bishop Conglomerate represents part of an integrated river system that was a precursor to the Platte River remains unresolved.

Extensional collapse of the eastern Uinta Mountains was marked by the cessation of Bishop Conglomerate fluvial deposition and the onset of Browns Park Formation sedimentation within the Browns Park graben beginning ca. 25 Ma. Tuffaceous sandstone and siltstone and minor quantities of carbonate accumulated in a mosaic of fluvial and lacustrine environments representing an internally drained basin. Detrital sanidine age and geochemical data for

young (younger than 40 Ma) grains also support a volcanic ash-fall origin. Some of the grains originated from the Southern Rocky Mountain volcanic field, and were reworked into Browns Park Formation deposits. New MDAs of Browns Park Formation sediments that unconformably overlie Neoproterozoic UMG rocks in westernmost Browns Park provide evidence for a younger (12–8 Ma) phase of extensional collapse of the eastern Uinta Mountains that was associated with 10–20 km of northwestward-directed lengthening of the Browns Park graben. These data are compatible with models for two stages of post-Laramide epeirogenic uplift of the Uinta Mountains region, including post-12 Ma tectonism that set the stage for subsequent integration of the Green and Colorado Rivers after 8 Ma.

INTRODUCTION

The northern boundary of the Colorado Plateau physiographic province (western USA) is the Precambrian-cored east-west-trending Uinta Mountains uplift that formed in the Laramide orogeny, ca. 70–50 Ma (Hansen, 1984, 1986; Bradley, 1995) (Fig. 1). Like the Kaibab uplift and Grand Canyon to the south, this uplift is a key physiographic barrier for understanding Cenozoic drainage evolution of the Colorado Plateau. Powell (1876) found it enigmatic that the Green River cut a deep gorge (Canyon of Lodore) orthogonal to the Uinta Mountains uplift, and postulated that the river course predated the uplift (antecedence). Sears (1924) postulated the opposite, that the uplift predated the Green River and that the river course was established at higher stratigraphic levels such that it maintained this path as the river incised resistant rocks in the core of the uplift (superposition). Since then, studies of the integration of the Green River and Colorado River systems have entertained more complex histories involving landscapes where modern river courses reflect linkage of internally drained basins and modification of paleoriver segments through a combination of piracy and downward integration (Blackwelder, 1934; Lucchitta, 1972; Pederson and Hadder, 2005; Aslan et al., 2014; Karlstrom et al., 2014; Kimbrough et al., 2015) coupled with Cenozoic uplift (Karlstrom et al., 2012).

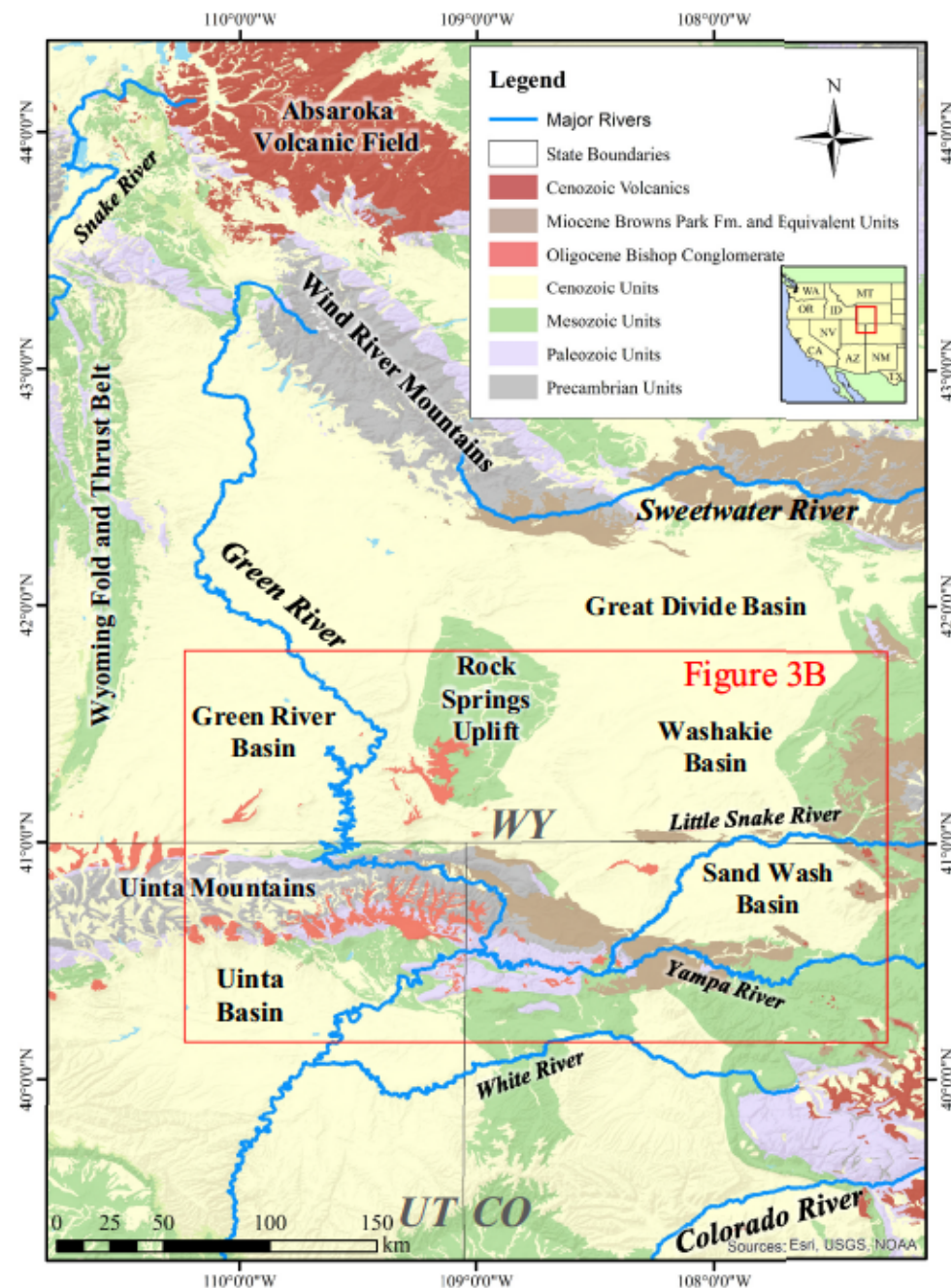


Figure 1. Map showing generalized geology of western Wyoming (WY), northwestern Colorado (CO), and northeastern Utah (UT). Red box outlines the general study area (Fig. 3B). Fm.—formation. Inset map shows the study area within the western United States.

Reconstructing Cenozoic landscapes is commonly hampered by imprecise chronologies that rely on techniques such as paleomagnetostigraphy, biostratigraphic correlation to land-mammal ages, and where available, radiometric dating of volcanic deposits. The application of detrital zircon U-Pb geochronology has provided another tool to constrain the ages of terrestrial clastic deposits (e.g., Stewart et al., 2001; Hodges et al., 2005; Gehrels et al., 2011; Fan et al., 2015), and newly emerging techniques such as detrital sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology offer potentially even more precise means for establishing temporal frameworks for interpreting the provenance of terrestrial sediments and landscape evolution (Hereford et al., 2016; Karlstrom et al., 2017). Within the western United States, Laramide intermontane basins are important archives

of Cenozoic landscape evolution (Lillgraven and Ostresh, 1988; Smith et al., 2003, 2008; McMillan et al., 2006), and these detrital dating techniques provide excellent opportunities to improve our understanding of paleogeography.

The Cenozoic geologic history of southwestern Wyoming, northeastern Utah, and northwestern Colorado (Fig. 1) provides the framework for studying the evolution of the upper Green River system beginning with Lakes Gosiute and Uinta during the Eocene (Bradley, 1936; Hansen, 1969a, 1969b, 1984, 1986; Smith et al., 2008) (Fig. 2A). As shown in Figure 2B, Hansen (1986) suggested that the Eocene lakes were replaced by an Oligocene eastward-flowing river system (ancestral Platte River) originating in the mountains of southwestern Wyoming and northeastern Utah. In this model, subsequent Miocene collapse

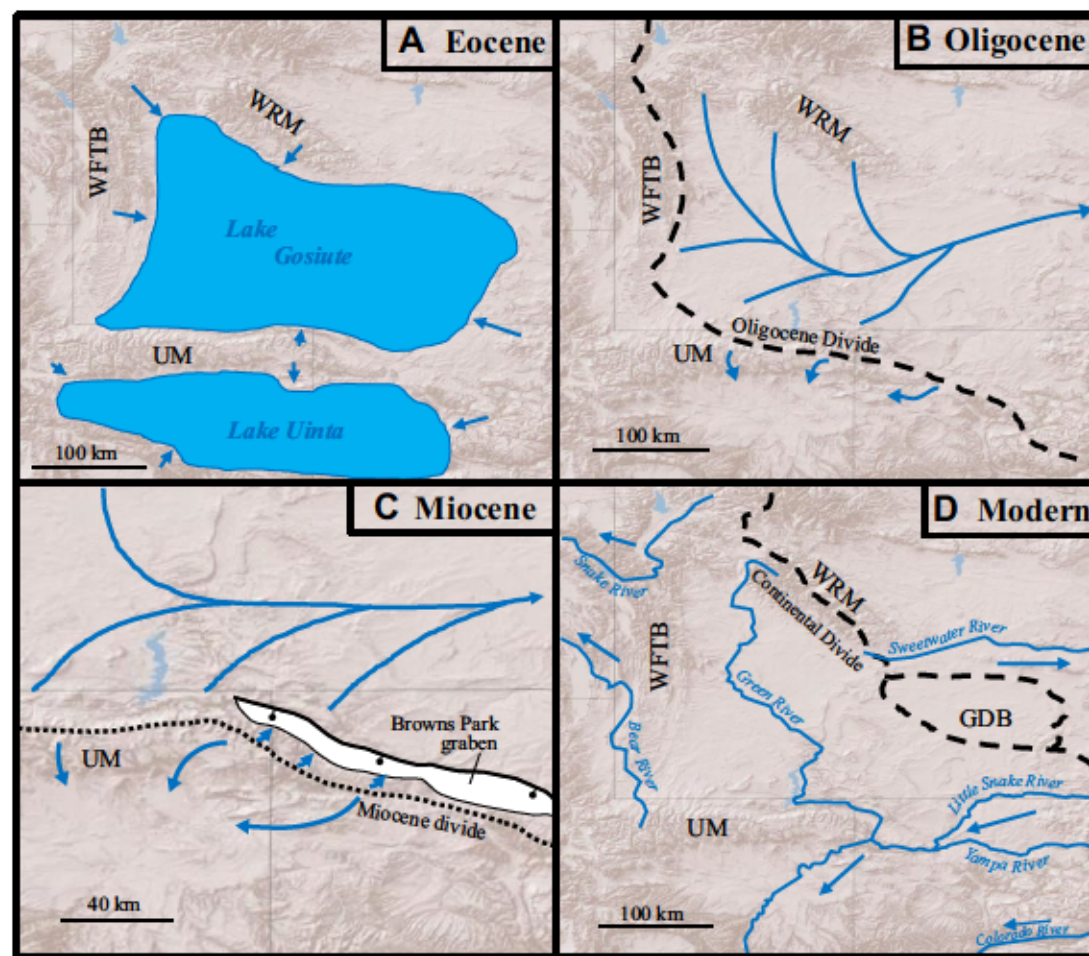


Figure 2. Paleogeographic maps showing Hansen's (1986) hypothesized evolution of Cenozoic fluvial systems in the Uinta Mountains region. (A) During the Eocene, internal drainage prevailed in the Rocky Mountain region and rivers drained into Lakes Gosiute and Uinta, located north and south of the Uinta Mountains, respectively. UM—Uinta Mountains; WFTB—Wyoming fold and thrust belt; WRM—Wind River Mountains. (B) During the Oligocene, an eastward-flowing ancestral Platte River system flowed through the Green River Basin. (C) During the Miocene, collapse of the eastern Uinta Mountains formed the Browns Park graben (note that the scale differs from that of A, B, and D). (D) From the Miocene to the present, drainage reorganization produced the modern river systems of the region. Blue arrows show flow direction of rivers. GDB—Great Divide Basin.

of the eastern Uinta Mountains and the formation of the Browns Park graben set the stage for drainage reorganization and capture of the ancestral eastward-flowing Platte River (Fig. 2C). Today the Green River flows southward across the Uinta Mountains through Lodore Canyon, and joins the Yampa River in Echo Canyon (Fig. 2D). Details of the transition from Eocene lakes to the modern Green River system, however, remain poorly understood. In particular, the ages of key Cenozoic deposits, the pathways of ancient rivers, and the precise timing of the collapse of the eastern Uinta Mountains remain poorly constrained (Hansen, 1986; Pederson and Haddad, 2005).

The purpose of this paper is to integrate sedimentologic and stratigraphic observations, provenance information, and age constraints provided by new detrital zircon and sanidine data to improve our understanding of Cenozoic landscape evolution in the Uinta Mountains region. Specifically, this study (1) provides new constraints on the timing of extensional collapse of the eastern Uinta Mountains, which set the stage for final integration of the Green River across the Uinta Mountains, (2) examines the Cenozoic fluvial record of the Uinta Mountains region and questions the interpretation of a late Eocene to Oligocene eastward-flowing ancestral Platte River system, and (3) highlights the prevalence of far-transported Oligocene–Miocene ash-fall grains and sediment reworking in the Bishop Conglomerate and Browns Park Formation, which has important implications for using detrital zircon and sanidine data for interpreting the provenance and depositional ages of clastic sediments that have accumulated during times of extensive volcanic activity.

■ GEOLOGIC SETTING

Late Cretaceous to early Cenozoic Laramide tectonism in southwestern Wyoming, northeastern Utah, and northwestern Colorado created uplifts in the study area, including the east-west-trending Uinta Mountains and the north-south-trending Rock Springs uplift (Fig. 1). North of the Uinta Mountains and west of the Rock Springs uplift, Eocene basin fill, including the Green River and Bridger Formations, accumulated in the southern Green River Basin (Bridger Basin) until ca. 47 Ma (Murphey and Evanoff, 2007; Smith et al., 2008) (Fig. 3A). South of the Uinta Mountains, Eocene sedimentary rocks represented by the Green River, Uinta, and Duchesne River Formations, filled the Uinta Basin until at least ca. 40 Ma (Mauger, 1977; Bryant et al., 1989; Prothero and Swisher, 1992; Sprinkel, 2015). Eocene deposits (Green River and Bridger Formations) filled the Washakie and Sand Wash Basins to the east and southeast of the Rock Springs uplift, respectively, until ca. 46–45 Ma (Smith et al., 2008).

A regional unconformity separates Eocene and older Laramide basin fill from younger units in Browns Park and the Green River and Uinta Basins (Fig. 3A). This unconformity is referred to as the Gilbert Peak erosion surface in the region (Bradley, 1936), and the Green Mountain erosion surface in the Sand Wash Basin (Buffler, 1967). This surface bevels rocks from the core of the Uinta Mountains (Neoproterozoic Uinta Mountain Group) to lower Paleogene units in the surrounding Laramide basins. In northeastern Utah, southwestern Wyoming, and parts of northwestern Colorado, this surface is overlain by the

upper Eocene–Oligocene Bishop Conglomerate. In some parts of northwestern Colorado, the erosion surface is overlain by the upper Oligocene–Miocene Browns Park Formation (Hansen, 1986; Buffler, 2003). Prior studies show that the Browns Park Formation locally overlies the Bishop Conglomerate along the eastern flank of the Uinta Mountains (Hansen, 1986).

The upper Eocene–Oligocene Bishop Conglomerate was first described by Powell (1876) and studied in detail by Bradley (1936) and Hansen (1984, 1986). The unit consists predominantly of fluvial sandy conglomerate, tuffaceous conglomeratic sandstone, and lesser amounts of tuffaceous mudrock that fill paleovalleys in northeastern Utah, southwestern Wyoming, and northwestern Colorado. In general, the Bishop Conglomerate is unlithified but erosionally resistant, and crops out on plateaus and mesa tops. Strata are typically flat lying, tens of meters to a few hundred meters thick, and unconformably overlie gently dipping Precambrian through early Paleogene rocks in both the Uinta and Green River Basins. The sediment source of the Bishop Conglomerate is interpreted to be primarily from the Uinta Mountains, because the deposits thin and clast sizes decrease radially away from the range front (Bradley, 1936; Hansen, 1986). Compositionally, clasts are dominated by red sandstones of the Neoproterozoic Uinta Mountain Group, and Paleozoic limestones and sandstones (Hansen, 1986). Bradley (1936) and Hansen (1986) interpreted the Bishop Conglomerate to represent a time of alluvial sedimentation associated with tectonic quiescence and climatic change from hot and moist conditions in the Eocene to cooler and drier conditions during the Oligocene. The age of the Bishop Conglomerate is constrained by $^{40}\text{Ar}/^{39}\text{Ar}$ ages (ca. 34–30 Ma) of volcanic tuffs located along the southern flank of the Uinta Mountains near Vernal, Utah (Kowallis et al., 2005). K–Ar ages of tuff of the Bishop Conglomerate located near Vernal suggest a slightly younger minimum age (ca. 29–26 Ma) (Damon, 1970; Hansen, 1986). The tuffs are thought to have been produced by caldera-forming eruptions within the Basin and Range extensional province in central Utah, including the Cottonwood caldera (Kowallis et al., 2005).

The upper Oligocene–Miocene Browns Park Formation was also first described by Powell (1876) and was studied in detail by Buffler (1967, 2003), Izett (1975), Hansen (1984, 1986), and Luft (1985) in northwestern Colorado and northeastern Utah. The Browns Park Formation contains tuffaceous sandstone and mudrock with minor amounts of limestone that represent a mosaic of fluvial, lacustrine, and eolian environments (Izett, 1975; Luft, 1985; Hansen, 1986; Buffler, 2003). Volcanic tuffs are locally abundant, and a conglomeratic facies underlies the fine-grained volcanoclastic deposits in the Sand Wash Basin near the Park Range and Flat Tops of Colorado (Kucera, 1962; Buffler, 2003). Although the eolian facies of the Browns Park Formation suggest that this unit covered a broad area originally (Izett, 1975), the thickest (to 500 m thick) and best preserved deposits are in extensional grabens such as Browns Park and the Yampa River valley south of Steamboat Springs in Colorado. Hansen (1986) documented that the collapse of the eastern Uinta Mountains at the end of the deposition of the Bishop Conglomerate provided the accommodation necessary to accumulate and preserve the Browns Park Formation. K–Ar ages of tuff from the Browns Park Formation range from ca. 25 to 8 Ma (Izett, 1975; Luft, 1985; Buffler, 2003).

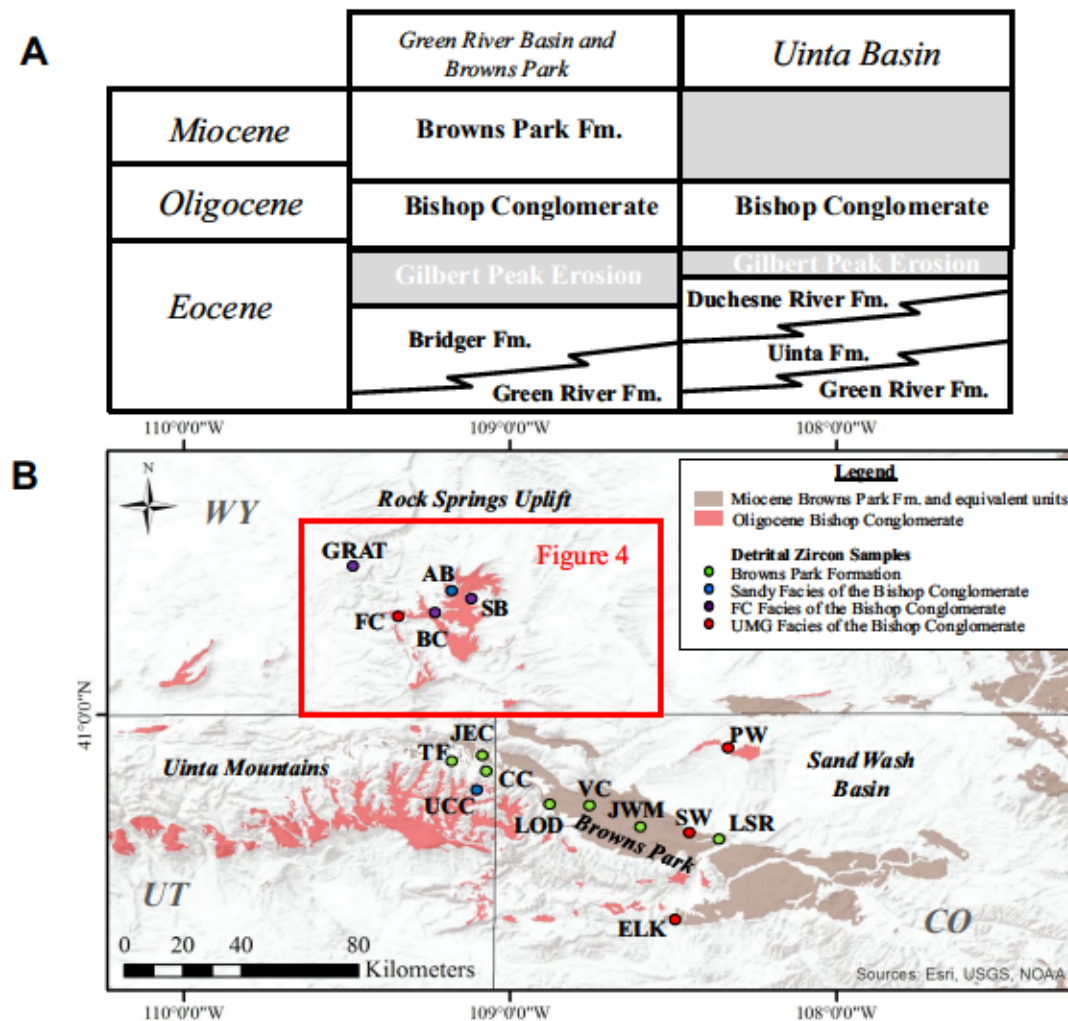


Figure 3. (A) Stratigraphic correlation chart of Eocene–Miocene units in the Green River and Uinta Basins and Browns Park area (modified from Hansen, 1984). Major unconformities are shown by gray fill pattern. Fm.—formation. (B) Map showing distribution of Oligocene Bishop Conglomerate and Miocene Browns Park Formation, and locations of detrital zircon samples. Location of Figure 4 is shown. Detrital zircon sample abbreviations: GRAT—Green River Airport Terrace; AB—Antelope Butte; BC—Bitter Creek; CC—Crouse Canyon; ELK—Elk Springs; FC—Firehole Canyon; JEC—Jesse Ewing Canyon; JWM—John Weller Mesa; LOD—Lodore Canyon; PW—Powder Wash; SW—Sand Wash; SB—South Baxter; TF—Taylor Flat; UCC—Upper Crouse Canyon. Note that location Firehole Canyon (FC) represents two samples; one sample of the UMG (Uinta Mountain Group) facies of the Bishop Conglomerate (shown as a red dot) and a second sample (not shown) representing the FC facies of the Bishop Conglomerate. Detrital sandstone samples were also acquired at BC, ELK, and JWM. Sources: Esri—<https://www.esri.com/>; USGS—U.S. Geological Survey; NOAA—U.S. National Oceanic and Atmospheric Administration. WY—Wyoming; UT—Utah; CO—Colorado.

METHODS

Field work consisted of describing and sampling selected outcrops of the Bishop Conglomerate and the Browns Park Formation in areas including (1) the southern flank of the Rock Springs uplift and the adjacent Green River Basin (Firehole Canyon, Aspen Mountain, Green River), (2) the Browns Park region of northeastern Utah and northwestern Colorado, and (3) parts of the western Sand Wash Basin of northwestern Colorado (Elk Springs, Powder Wash) (Fig. 3B). We also described clast types at selected localities. Paleocurrent

measurements were limited to the Firehole Canyon area where Bishop Conglomerate is locally cemented. In many areas, the Bishop Conglomerate is un lithified. Paleocurrents were primarily measured using imbricated clasts within conglomeratic units, although in a few instances measurements were made on foresets of trough cross-bedded sandstones. Measurements were made at a total of 18 stations, and 5–8 measurements were recorded at each station.

We collected 17 detrital zircon and 3 detrital sandstone samples from gravelly, sandy, and tuffaceous beds of the Bishop Conglomerate and Browns Park Formation (Supplemental Item 1[†]). Both zircon and K-feldspar grains were

Supplemental Item 1. Detailed Bishop Conglomerate and Browns Park Formation detrital zircon and sandstone sample locality information. Please visit <http://doi.org/10.1130/GES01523.S1> or the full-text article on www.gsapubs.org to view Supplemental Item 1.

Sample	Location	Sample #	Latitude	Longitude	Sample Type	Sample Age (Ma)	Sample Age (Ma)
GRAT-1	Green River Airport Terrace	GRAT-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
GRAT-2	Green River Airport Terrace	GRAT-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
AB-1	Antelope Butte	AB-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
AB-2	Antelope Butte	AB-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
BC-1	Bitter Creek	BC-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
BC-2	Bitter Creek	BC-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
SB-1	South Baxter	SB-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
SB-2	South Baxter	SB-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
JEC-1	Jesse Ewing Canyon	JEC-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
JEC-2	Jesse Ewing Canyon	JEC-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
TF-1	Taylor Flat	TF-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
TF-2	Taylor Flat	TF-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
UCC-1	Upper Crouse Canyon	UCC-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
UCC-2	Upper Crouse Canyon	UCC-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
CC-1	Crouse Canyon	CC-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
CC-2	Crouse Canyon	CC-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
VC-1	Vermilion Canyon	VC-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
VC-2	Vermilion Canyon	VC-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
JWM-1	John Weller Mesa	JWM-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
JWM-2	John Weller Mesa	JWM-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
SW-1	Sand Wash	SW-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
SW-2	Sand Wash	SW-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
LSR-1	Lodore Canyon	LSR-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
LSR-2	Lodore Canyon	LSR-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
ELK-1	Elk Springs	ELK-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
ELK-2	Elk Springs	ELK-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
PW-1	Powder Wash	PW-1	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5
PW-2	Powder Wash	PW-2	40°55'N	108°55'W	FC Facies of the Bishop Conglomerate	5.5	5.5

[†]Supplemental Item 1. Detailed Bishop Conglomerate and Browns Park Formation detrital zircon and sandstone sample locality information. Please visit <http://doi.org/10.1130/GES01523.S1> or the full-text article on www.gsapubs.org to view Supplemental Item 1.

Supplemental Item 2. Summary of Bishop Conglomerate and Browns Park Formation detrital zircon U-Pb data. Please visit <http://doi.org/10.1130/GES01523.S2> or the full-text article on www.gsapubs.org to view Supplemental Item 2.

Sample	Grain #	Age (Ma)	Uncertainty (Ma)	Discordance (%)	Reverse (%)
Bishop Conglomerate	1	1000	±10	0	0
	2	1000	±10	0	0
	3	1000	±10	0	0
	4	1000	±10	0	0
	5	1000	±10	0	0
	6	1000	±10	0	0
	7	1000	±10	0	0
	8	1000	±10	0	0
	9	1000	±10	0	0
	10	1000	±10	0	0
Browns Park Formation	11	1000	±10	0	0
	12	1000	±10	0	0
	13	1000	±10	0	0
	14	1000	±10	0	0
	15	1000	±10	0	0
	16	1000	±10	0	0
	17	1000	±10	0	0
	18	1000	±10	0	0
	19	1000	±10	0	0
	20	1000	±10	0	0

Supplemental Item 2. Summary of Bishop Conglomerate and Browns Park Formation detrital zircon U-Pb data. Please visit <http://doi.org/10.1130/GES01523.S2> or the full-text article on www.gsapubs.org to view Supplemental Item 2.

DS3. $^{40}\text{Ar}/^{39}\text{Ar}$ data repository for:

Cenozoic collapse of the eastern Uinta Mountains and drainage evolution of the Uinta Mountain region

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Supplemental Item 3. Summary of Bishop Conglomerate and Browns Park Formation detrital sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ data. Please visit <http://doi.org/10.1130/GES01523.S3> or the full-text article on www.gsapubs.org to view Supplemental Item 3.

DB4. Maximum depositional age calculation data repository for:

Cenozoic collapse of the eastern Uinta Mountains and drainage evolution of the Uinta Mountain region

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Supplemental Item 4. Summary of maximum depositional age calculations for Bishop Conglomerate and Browns Park Formation samples using both detrital zircon U-Pb and sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ data. Please visit <http://doi.org/10.1130/GES01523.S4> or the full-text article on www.gsapubs.org to view Supplemental Item 4.

separated from bulk samples using standard methods. Detrital zircon (DZ) U-Pb ages of ~100 detrital grains from each sample were determined by laser ablation-multicollector-inductively coupled plasma-mass spectrometry (LA-MC-ICP-MS) at the Arizona LaserChron Center (Tucson, Arizona). For individual zircon grains older than 900 Ma, $^{207}\text{Pb}/^{206}\text{Pb}$ ages are reported, whereas for zircon grains younger than 900 Ma we report the $^{206}\text{Pb}/^{238}\text{U}$ ages. Zircon grain ages were filtered using a 20% discordance filter and a 5% reverse discordance filter. Complete details are tabulated in Supplemental Item 2².

Detrital sanidine (DS) geochronology involved hand-picking clear K-feldspar grains in an attempt to concentrate sanidine for $^{40}\text{Ar}/^{39}\text{Ar}$ dating of individual crystals (~45–65 grains per sample) via single-crystal laser fusion. Analyses were completed at the New Mexico Geochronology Research Laboratory (NMGRRL) using an ARGUS VI multicollector mass spectrometer, and individual grains yielded precision of ~0.05%–0.2% (2 standard deviations) for 35–20 Ma with variation dependent upon grain size and radiogenic yield. Based on age and K/Ca value, this method was very successful for Browns Park sample Tbp 7912-2 and less so for the Bishop Conglomerate sample BC71112-1 that yielded several microcline grains older than 500 Ma. Complete details are tabulated in Supplemental Item 3³. DS age spectra were compared to ignimbrite sanidine data from the NMGRRL database in an effort to determine source area.

Zircon and sanidine data sets were used to assess the maximum depositional ages (MDAs) of clastic deposits. Clastic deposits typically represent reworking of material that was previously crystallized or formed. Sometimes billions of years separate the formation of a parent rock and deposition of its clastic residuum. As a result, radiometric ages of constituent minerals may not provide the best estimate of depositional age, but the deposit can be no older than the youngest constituent minerals. This gap between the age of a deposit and the MDA given by the youngest crystallized minerals in a deposit is referred to as its lag time (e.g., Cervený et al., 1988; Thomas et al., 2004). This lag time can be minimized, and the MDA may approximate actual depositional ages if fluvial sediments accumulate during times of semicontinuous felsic volcanic activity (e.g., Fan et al., 2015).

Detrital mineral ages provide an estimate of the MDA of a clastic deposit (e.g., Dickinson and Gehrels, 2009; May et al., 2013), assuming that it did not undergo temperatures that would have enabled thermal diffusion of radiogenic daughter product. For zircon, diffusive loss of daughter product occurs above temperatures of ~800–1000 °C (Cherniak, 2010), and for sanidine, it is at least 250 °C (McDougall and Harrison, 1999). While it may seem straightforward to identify the youngest age, the uncertainty (often attributable to random error or detection limits) surrounding any individual analysis may complicate a decision. In addition, young zircons (younger than 100 Ma) may carry a larger uncertainty (as a percentage of age) because it becomes increasingly difficult to distinguish the radiogenic ^{206}Pb component from the common background Pb. Complicating matters, the concentration of radiogenic Pb (^{206}Pb) depends on the concentration of ^{238}U , so some zircons should ostensibly have higher precision than others, and those analyses should be considered more robust than others. Because of magma system complexities and zoning, a collection

of U-Pb ages of zircons collected from a single volcanic rock may show a distribution with overlapping uncertainties (typically 1 σ with LA-ICP-MS data sets) such that the youngest age distribution in a sample may be a better estimate of the age of a volcanic event than the youngest grain.

As outlined by Dickinson and Gehrels (2009), there are at least five primary methods that can be used as a basis for determining MDAs: (1) the youngest single grain age within a distribution of detrital ages, (2) the youngest probability plot (YPP) within a probability density function of a sample of detrital ages, (3) the weighted mean age of the youngest cluster of 2 or more grain ages with 1 σ overlap of uncertainty [YC 1 σ (2+)], (4) the weighted mean age of the youngest cluster of 3 or more grain ages overlapping at 2 σ [YC 2 σ (3+)], and (5) a Monte Carlo analysis of the uncertainty of the youngest subset of detrital ages in an age distribution. This approach is incorporated in the Excel Isoplot add-in (Ludwig, 2012). This approach is similar to the YPP determination, but the actual uncertainty around this youngest peak is explored differently. In this study, MDAs of both zircon and sanidine samples were calculated as the mean standard weighted ages of the youngest group of three or more detrital grains from each sample [YC 2 σ (3+) of Dickinson and Gehrels, 2009] using Isoplot (Ludwig, 2012) (Table 1). The details of these calculations are presented in Supplemental Item 4⁴.

The mean square of weighted deviates (MSWD) is commonly used to assess how well a cluster of ages may represent a single population (Wendt and Carl, 1991). In Isoplot (Ludwig, 2012), the weighted mean (WM) is calculated from data for a series of grains by weighting each grain according to their variance with small uncertainties more heavily weighted than those with large uncertainties. The MSWD provides a measure of the deviation of the data from a normal distribution with respect to analytical error. Ideally an MSWD of 1 indicates that data are scattered about a mean value in accordance with the analytical error. An MSWD that is >1 indicates that the age is likely unreliable, signifying a high degree of scatter not accounted for by the analytical uncertainty alone. Alternatively, an MSWD <1 suggests that the errors assigned to the analysis have been overestimated, but the mean age may be considered reliable.

BISHOP CONGLOMERATE

The Bishop Conglomerate was studied in detail north of the Uinta Mountains in southwestern Wyoming and along the perimeter of the outcrop belt of the Browns Park Formation in northeastern Utah and northwestern Colorado (Fig. 3B). This unit unconformably overlies Eocene and older units in these areas. The majority of the outcrops are represented by poorly consolidated conglomeratic sandstone and sandy conglomerate with boulder- to pebble-sized clasts dominated by red sandstone and light gray limestone and sandstone derived from the Uinta Mountains, similar to the composition of the extensive outcrops of Bishop Conglomerate on the south flanks of the Uinta Mountains (Hansen, 1986; Kowallis et al., 2005). Specifically, the red sandstones are reworked from

TABLE 1. MAXIMUM DEPOSITIONAL AGE CALCULATIONS FOR BISHOP CONGLOMERATE AND BROWNS PARK FORMATION DETRITAL ZIRCON AND SANIDINE DATA

Sample location and abbreviation*	Unit [†]	Facies [‡]	Lithology	Detrital zircon MDA** (Ma)	MSWD* for zircon MDA	Youngest zircon U-Pb age (Ma)	Detrital sanidine MDA** (Ma)	Youngest sanidine ⁴⁰ Ar/ ³⁹ Ar age (Ma)	MSWD for sanidine MDA
Firehole Canyon, WY (FC)	Tbc	UMG	sandstone	34.7 ± 4.7 (n = 3)	12	33.4 ± 0.9			
Powder Wash, CO (PW)	Tbc	UMG	sandstone	32.2 ± 4.4 (n = 3)	9.8	29.8 ± 1.3			
Elk Springs, CO (ELK)	Tbc	UMG	conglomerate	29.8 ± 0.9 (n = 12)	0.46	29.4 ± 1.2	34.9 ± 0.14 (n = 9)	30.6 ± 0.11	11.5
Sand Wash, CO (SW)	Tbc	UMG	sandstone	27.1 ± 1.2 (n = 2)	0.35	26.8 ± 1.7			
Green River, WY (GRAT)	Tbc	FC	conglomerate	35.5 ± 0.6 (n = 22)	0.5	30.9 ± 5.9			
Firehole Canyon, WY (FC)	Tbc	FC	sandstone	33.4 ± 0.3 (n = 26)	0.61	31.6 ± 1.9			
Bitter Creek, WY (BC)	Tbc	FC	conglomerate	34.0 ± 1.8 (n = 6)	6	32.5 ± 0.9	33.4 ± 0.16 (n = 8)	32.3 ± 0.07	20
South Baxter, WY (SB)	Tbc	FC	conglomerate	33.9 ± 0.6 (n = 10)	4.5	31.3 ± 1.9			
Antelope Butte, WY (AB)	Tbc	SAND	tuff	31.2 ± 0.6 (n = 25)	3	30.0 ± 0.7			
Upper Crouse Canyon, UT (UCC)	Tbc	SAND	sandstone	27.0 ± 0.3 (n = 13)	1.1	24.3 ± 5.0			
Taylor Flat, UT (TF)	Tbp		sandstone	8.4 ± 0.3 (n = 16)	0.63	7.8 ± 0.7			
Jesse Ewing Canyon, UT (JEC)	Tbp		tuff	9.2 ± 1.4 (n = 3)	0.79	8.4 ± 2.0			
Crouse Canyon, UT (CC)	Tbp		tuff	8.9 ± 0.6 (n = 17)	0.41	6.4 ± 4.5			
Lodore Canyon, CO (LC)	Tbp		sandstone	18.5 ± 1.8 (n = 4)	1.5	17.8 ± 1.3			
Vermillion Creek, CO (VC)	Tbp		tuff	13.7 ± 2.9 (n = 4)	0.37	12.7 ± 7.0			
John Weller Mesa, CO (JWM)	Tbp		sandstone	30.7 ± 1.7 (n = 9)	4.1	18.8 ± 0.7	27.5 ± 0.31 (n = 8)	16.6 ± 0.05	27
Little Snake River (LSR)	Tbp		tuff	24.5 ± 1.1 (n = 4)	0.19	24.1 ± 3.3			

*Abbreviations: WY—Wyoming; CO—Colorado; UT—Utah (see Fig. 3B for locations; detailed locations are reported in Supplemental Item 1 [see text footnote 1]). MSWD—mean square of weighted deviates.

[†]Tbc—Bishop Conglomerate, Tbp—Browns Park Formation.

[‡]UMG—Uinta Mountain Group facies, FC—Firehole Canyon facies, SAND—sandy facies.

**MDA—maximum depositional age based on weighted average tool in Isoplot (Ludwig, 2012). Uncertainties are reported at two standard deviations (2σ); n—number of detrital zircon grains used to calculate the maximum depositional age.

the Neoproterozoic Uinta Mountain Group (UMG) and light gray limestones and sandstones are derived from Paleozoic carbonates and siliciclastic units.

We identified and described three facies that compose the Bishop Conglomerate: (1) UMG facies, (2) Firehole Canyon (FC) facies, and (3) sandy facies. The UMG and FC facies locally interfinger in southwestern Wyoming, and the sandy facies, which is rare due to its friable texture, overlies older UMG facies deposits.

UMG Facies Sedimentology and Composition

The UMG facies of the Bishop Conglomerate is the predominant type within the southern Green River Basin region, armoring the north-sloping surfaces of Miller, Little, and Pine Mountains (Fig. 4). Overall, this facies represents a series of lenses of northward-thinning coarse-grained wedges that unconformably overlie Late Cretaceous–Eocene rocks of the Green River Basin and the adjacent Rock Springs uplift (Fig. 5A). The UMG facies also crops out over areas of the western Sand Wash Basin (Powder Wash), the eastern Uinta Mountains (Elk Springs), and within Browns Park (Sand Wash), west of the Little Snake River (Fig. 4). This facies is also at Little Mountain (Utah) and along parts of the south flank of the Uinta Mountains (Hansen, 1986).

Outcrops of the UMG facies range from ~3 to 30 m in thickness and are dominated by moderately to poorly sorted, boulder- to pebble-sized clasts of subangular to subrounded reddish sandstone and light gray limestone with

minor quantities of quartzite and chert clasts (Fig. 5B). Where cemented, the Bishop Conglomerate is crudely bedded, and contains lenses of cross-stratified sandstone. Paleocurrents measured from imbricated gravel clasts and trough cross-strata at Firehole Canyon located north of Miller Mountain show a predominant northeastward paleoflow direction (Fig. 4). Where exposed, the basal contact of the UMG facies is strongly erosional with several meters of local relief, and basal elevations of the conglomerate decrease to the north by ~600 m over a distance of ~35 km (average gradient of ~1°).

FC Facies Sedimentology and Composition

The FC facies of the Bishop Conglomerate is a newly recognized, rare but paleogeographically significant facies of the southern Green River Basin region. The facies is represented by a roughly east-west-trending outcrop belt located in the saddle between Aspen and Miller Mountains, and extends as far west as Green River, Wyoming (Fig. 4). Similar to the UMG facies, FC facies strata unconformably overlie Late Cretaceous–Eocene rocks. The best exposures of the FC facies are at Firehole Canyon, the saddle between Miller and Aspen Mountains, and on the bluffs south of Green River, Wyoming. Outside of the southern Green River Basin, FC facies outcrops are also present near Powder Wash. At this location, northeastward-thinning UMG facies pass laterally (eastward) into FC facies (Fig. 3B).

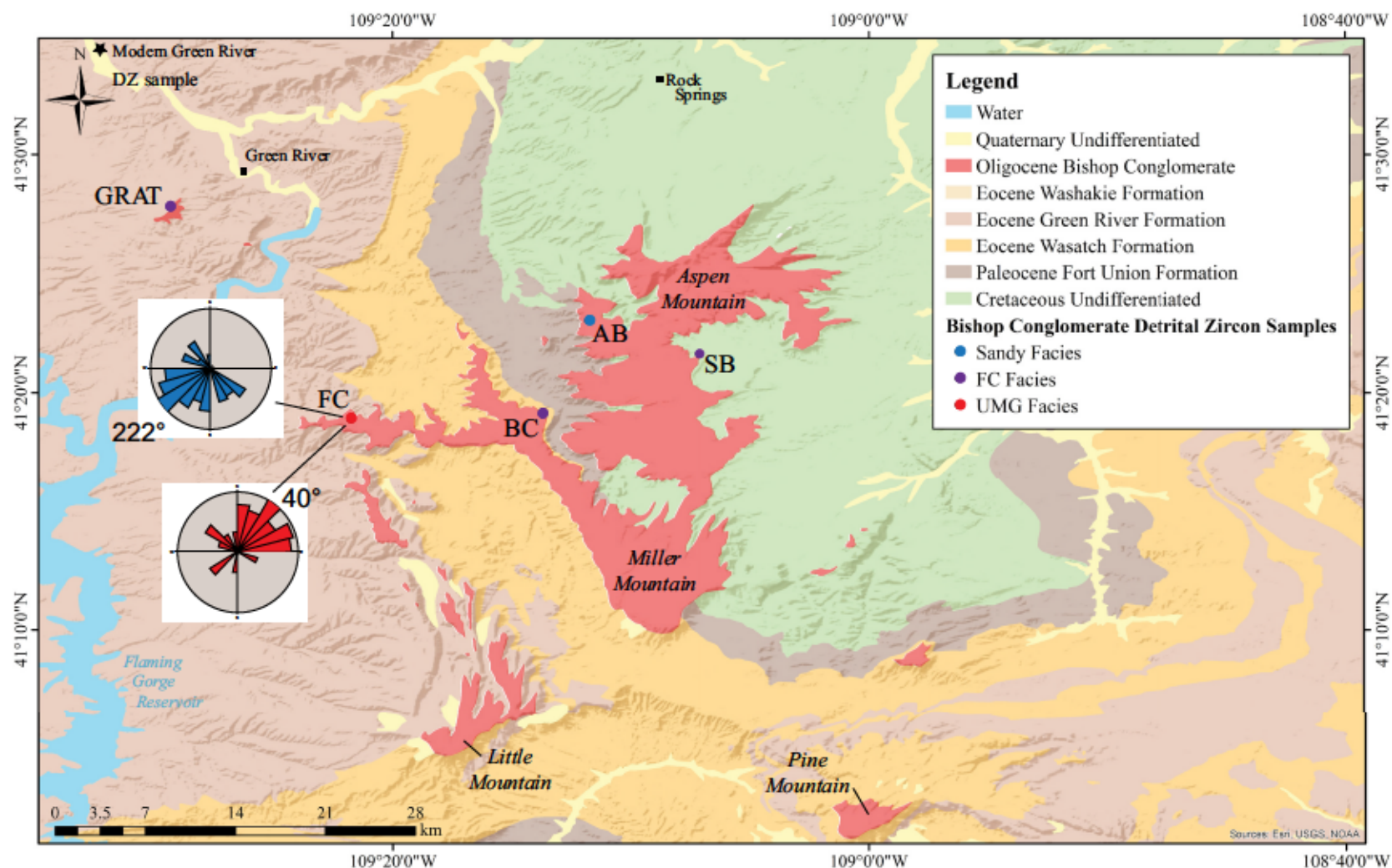


Figure 4. Geologic map of the southern Rock Springs uplift in southwestern Wyoming. Locations of paleocurrent data for the UMG (Uinta Mountain Group) and FC (Firehole Canyon) facies of the Bishop Conglomerate are shown. Detrital zircon sample and FC and UMG abbreviations are as in Figure 3B. Location of modern Green River detrital zircon sample (black star) is also shown.

Sedimentary rocks of the FC facies range from ~2 to 6 m in thickness and are dominated by moderately sorted, pebble- to cobble-sized, rounded to subrounded conglomerate with minor sandstone lenses. In addition to the smaller caliber of the clasts and the greater percentage of rounded clasts, the FC facies differs markedly in composition from the UMG facies. FC facies consist almost exclusively of black, gray, brown, and green quartzite

clasts (Fig. 5C). A small but significant quantity of aphanitic to porphyritic light gray, gray, and dark green andesitic pebbles are also present in the FC facies, but these types of volcanic clasts are completely absent from the UMG facies.

Carbonate-cemented conglomerate beds of the FC facies near Firehole Canyon consist of crudely-bedded conglomerate with rare lenses of trough

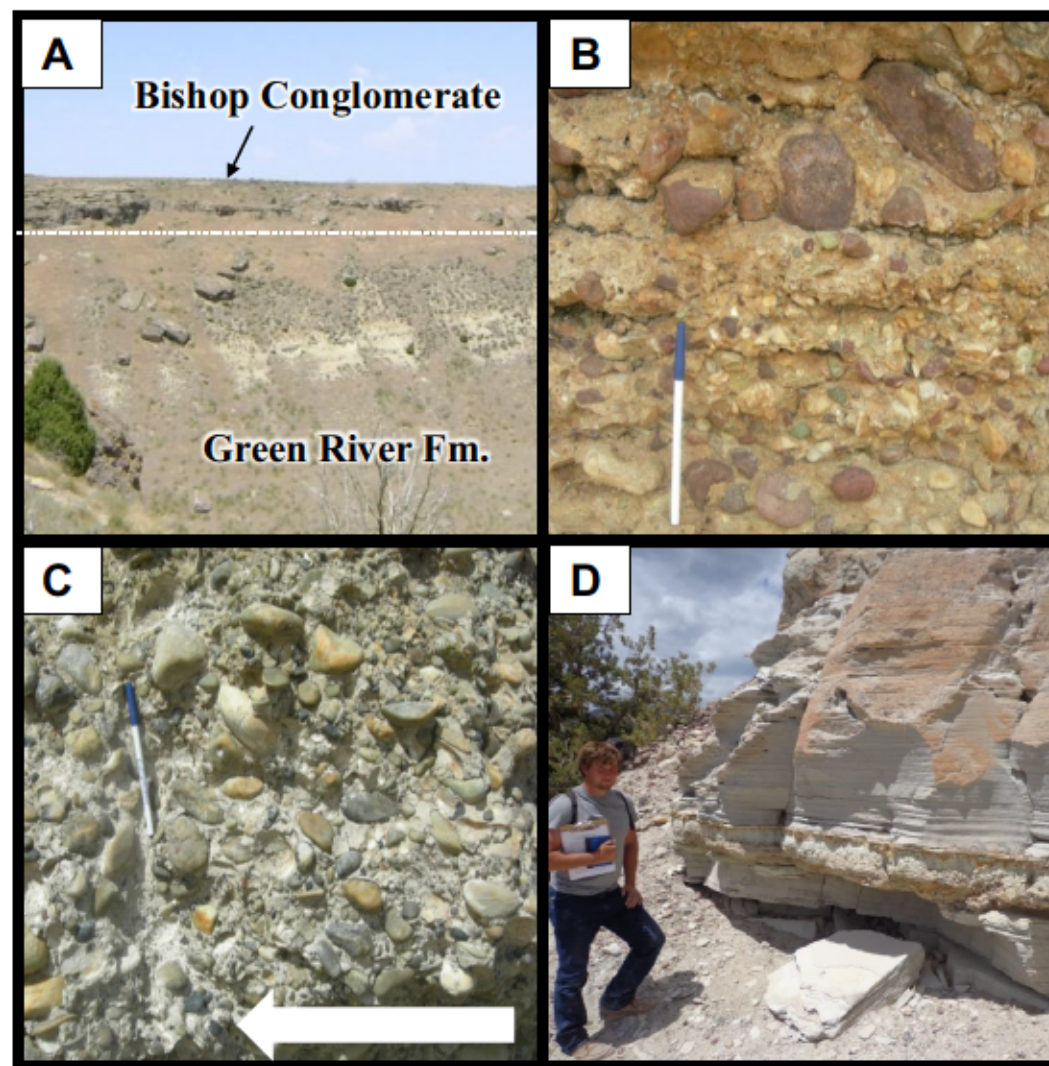


Figure 5. Bishop Conglomerate and Browns Park Formation field photographs from southwestern Wyoming and northeastern Utah. (A) Well-cemented, horizontally bedded Bishop Conglomerate overlies gently dipping strata of the Eocene Green River Formation at Firehole Canyon (location FC in Fig. 4). Thickness of the Bishop Conglomerate outcrop is ~5 m. (B) Outcrop photograph of UMG (Uinta Mountain Group) facies at Firehole Canyon showing abundant reddish sandstone cobbles eroded from outcrops of the Neoproterozoic Uinta Mountain Group (pen for scale). (C) Outcrop photograph of FC facies at Firehole Canyon. Clasts are chiefly rounded quartzite and imbrication shows paleoflow direction (white arrow) from right to left in the photograph (pen for scale). (D) Thinly bedded to laminated vitric tuff in the Taylor Flat area of western Browns Park (location TF in Fig. 3B). Bedding indicates that the tuff is reworked and probably accumulated in standing water (person for scale).

cross-bedded sandstone. The sandstone lenses are distinct channel-form bodies that are 2–3 m thick. Paleocurrents measured from imbricated gravels and trough cross-strata at Firehole Canyon show a predominant southwestward paleoflow direction, opposite to that of the UMG facies paleocurrents (Fig. 4). The basal contact of the FC facies is strongly erosional with several meters of local relief, and basal elevations of this contact decrease to the west by ~90 m over a distance of ~30 km.

Sandy Facies Sedimentology and Composition

North of the Uinta Mountains, sandy facies of the Bishop Conglomerate are relatively rare and poorly exposed. Where present, the sandy facies generally overlie the UMG facies. In the southern Green River Basin, the sandy facies are best exposed at Antelope Butte, which is located on the southwest flank of Aspen Mountain (Fig. 4). At Antelope Butte, as much as 60 m of inter-

bedded tuffaceous siltstone and sandstone and minor interbeds of siliceous limestone overlie the bouldery Bishop Conglomerate (Love and Blackmon, 1962; Kirschbaum, 1986). Love and Blackmon (1962) mapped these deposits as part of the Bishop Conglomerate, but were uncertain about this assignment given the similarity between these deposits and tuffaceous sandstones of the Browns Park Formation.

The other set of exposures of the sandy facies is in the eastern Uinta Mountains. In upper Crouse Canyon (UCC of Fig. 3B), mapping (by Sprinkel, 2006) showed that south-trending Bishop Conglomerate paleovalleys are filled, in part, by tuffaceous, locally cross-bedded, friable sandstone and siltstone. These beds overlie Neoproterozoic through Eocene bedrock and although poorly exposed, superficially resemble tuffaceous sedimentary rocks of the Browns Park Formation located ~10 km northeast in Browns Park. The sandy facies is also associated with classic outcrops of the Bishop Conglomerate on the Diamond Plateau near Vernal, Utah.

■ BROWNS PARK FORMATION

In the study area, deposits of the Browns Park Formation as much as 500 m thick primarily crop out in the Browns Park syncline, along the eastern margin of the Uinta Mountains, and in the southern Sand Wash Basin (Fig. 3B). The best exposures crop out in the Browns Park syncline, which represents a graben localized over the northern edge of the eastern Uinta Mountains range front, presumably related to extensional collapse along preexisting reverse faults (Hansen, 1986; Stone, 1993; Bradley, 1995; Sprinkel, 2006). In the Browns Park syncline, the Browns Park Formation consists chiefly of gray, interbedded, tuffaceous sandstone, siltstone, and lenses of light gray, vitric tuff (Izett, 1975; Honey and Izett, 1980; Luft, 1985) (Fig. 5D). Along the eastern margin of Browns Park between Vermillion Creek and the Little Snake River, the Browns Park Formation conformably overlies the Bishop Conglomerate, and both units dip steeply toward the axis of the Browns Park syncline. Elsewhere, near the western margin of Browns Park, Browns Park strata are relatively flat lying, and onlap gently dipping Neoproterozoic UMG sandstones. Zircon fission track ages from tuffaceous material indicate that the Browns Park Formation accumulated from ca. 25 to 8 Ma in the Browns Park syncline (Izett et al., 1970; Izett, 1975; Luft, 1985).

The lenticular geometry and the presence of well-preserved bedding indicate that Browns Park Formation tuffs are reworked ash fall that probably accumulated in shallow ponds (Hansen, 1986). Lenses of micritic, siliceous limestone, which are also interpreted as shallow ponds, are present locally and interfinger with tuffaceous units. Conglomeratic beds are 1–5 m thick and consist of clasts of locally derived reddish sandstone of the Uinta Mountain Group. The conglomeratic beds form discontinuous clastic wedges that interfinger with sandy and silty facies along the northern margins of westernmost Browns Park. Collectively, these deposits represent a mosaic of lacustrine and fluvial environments that are consistent with an internally drained

depositional system. The transition from fluvial coarse-grained facies of the Bishop Conglomerate to finer grained facies of the Browns Park Formation has been interpreted to reflect collapse of the eastern Uinta Mountains and formation of the Browns Park syncline (Hansen, 1986). Intraformational unconformities and faults that cut the Browns Park Formation show that deformation continued during and after the time represented by the Browns Park Formation.

■ DETRITAL ZIRCON AND SANIDINE GEOCHRONOLOGY

Detrital zircon and sanidine data were used to (1) further characterize the different facies of the Bishop Conglomerate, (2) evaluate the provenance of young (younger than 40 Ma) detrital grains in the Bishop Conglomerate and Browns Park Formation, (3) compare detrital zircon and sanidine results, and (4) constrain the timing of key events using MDAs of zircon and sanidine samples.

Bishop Conglomerate Facies

Comparison of the detrital zircon data of the UMG and FC facies clearly shows that the provenance of the two facies differs (Fig. 6A). While Archean (2700–2600 Ma), Paleoproterozoic (1850–1750 Ma), Mesoproterozoic (ca. 1450 Ma) and Grenville orogeny (1200–1000 Ma) detrital zircon U-Pb ages are well represented in both facies, detrital zircon grains of Grenville orogeny age are more abundant (39% of the grains analyzed) in the UMG facies compared to the FC facies sample. Another noteworthy difference is the Archean grain ages (Figs. 7A, 7B). The FC facies contains a ca. 2650 Ma age mode that is absent from the UMG facies. The FC facies sample also has Early Cretaceous (107–90 Ma) detrital zircon grains that are absent from the UMG facies sample (Fig. 6B). The most striking difference, however, is represented by the overwhelming quantity of grains younger than 80 Ma in the FC facies (47%) compared to the UMG facies (11%) (Fig. 6). The majority of these young grains are younger than 40 Ma. These differences in the U-Pb age populations of the UMG and FC facies are further confirmed by a Kolmogorov-Smirnov (K-S) test that shows that the two facies are statistically different ($p = 0.00$).

Browns Park Formation

Detrital zircon and sanidine ages for grains younger than 60 Ma in the Browns Park Formation samples range from 57 to 8 Ma, and all samples show major U-Pb age modes between 39 and 24 Ma (Figs. 8A–8E). When arranged in stratigraphic order, stratigraphically younger Browns Park Formation samples show successively younger U-Pb age distributions.

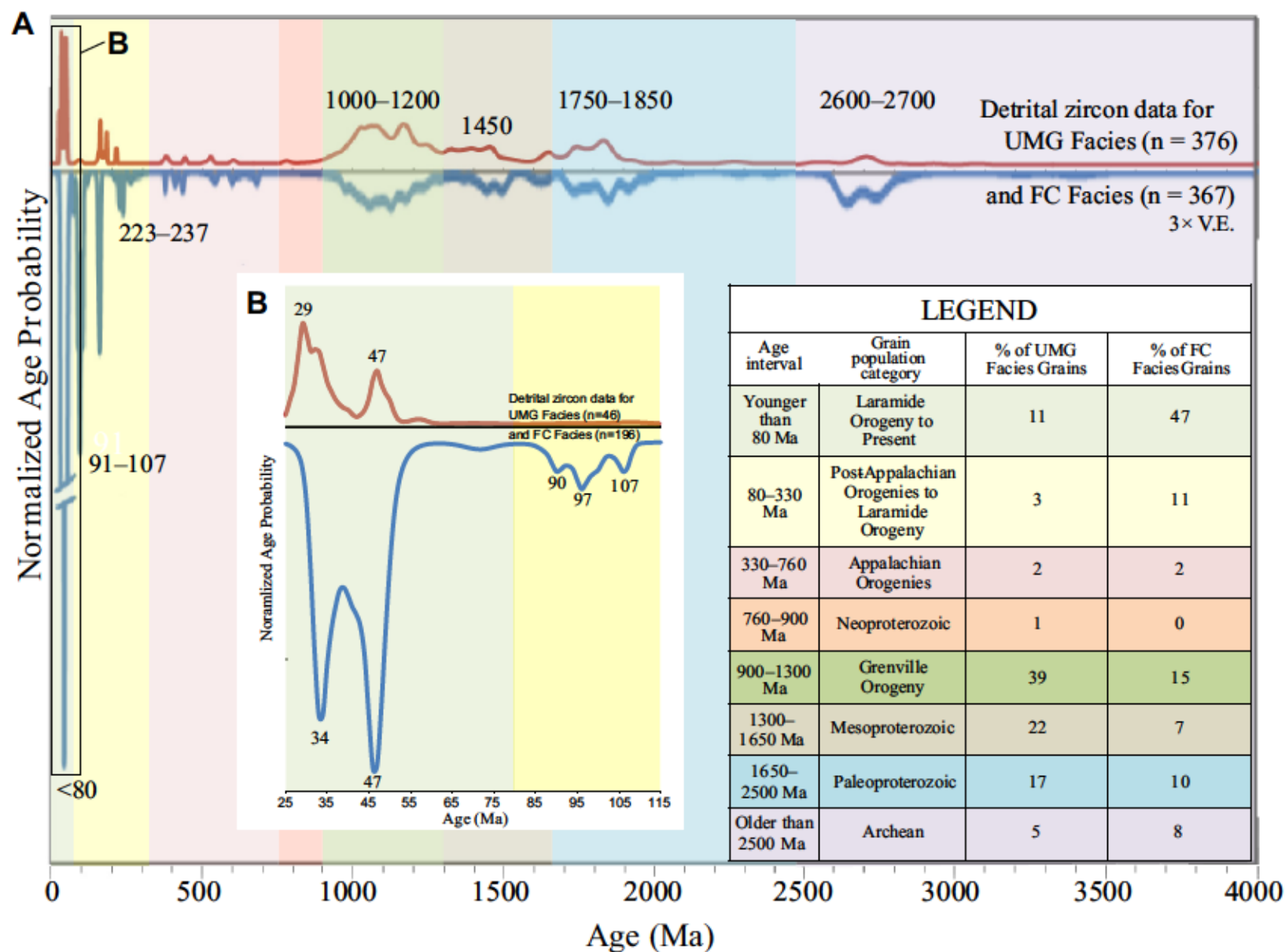


Figure 6. (A) Composite age probability curves comparing detrital zircon U-Pb data for UMG (Uinta Mountain Group; red curve) and FC (Firehole Canyon; blue curve) facies of the Bishop Conglomerate. Each composite curve represents data from four separate sample sites. The UMG curve includes data from samples FC, AB, PC, and ELK (locations shown in Fig. 3B). The FC curve includes data from samples GRAT, BC, SB, and FC (locations shown in Fig. 3B). Abbreviations as in Figure 2. (B) Detrital zircon U-Pb data for Cenozoic and Cretaceous (younger than 115 Ma) grains. Numbers in bold represent age ranges of detrital zircon U-Pb age distributions in Ma; n = total number of grains analyzed, and vertical exaggeration of both curves is 3x.

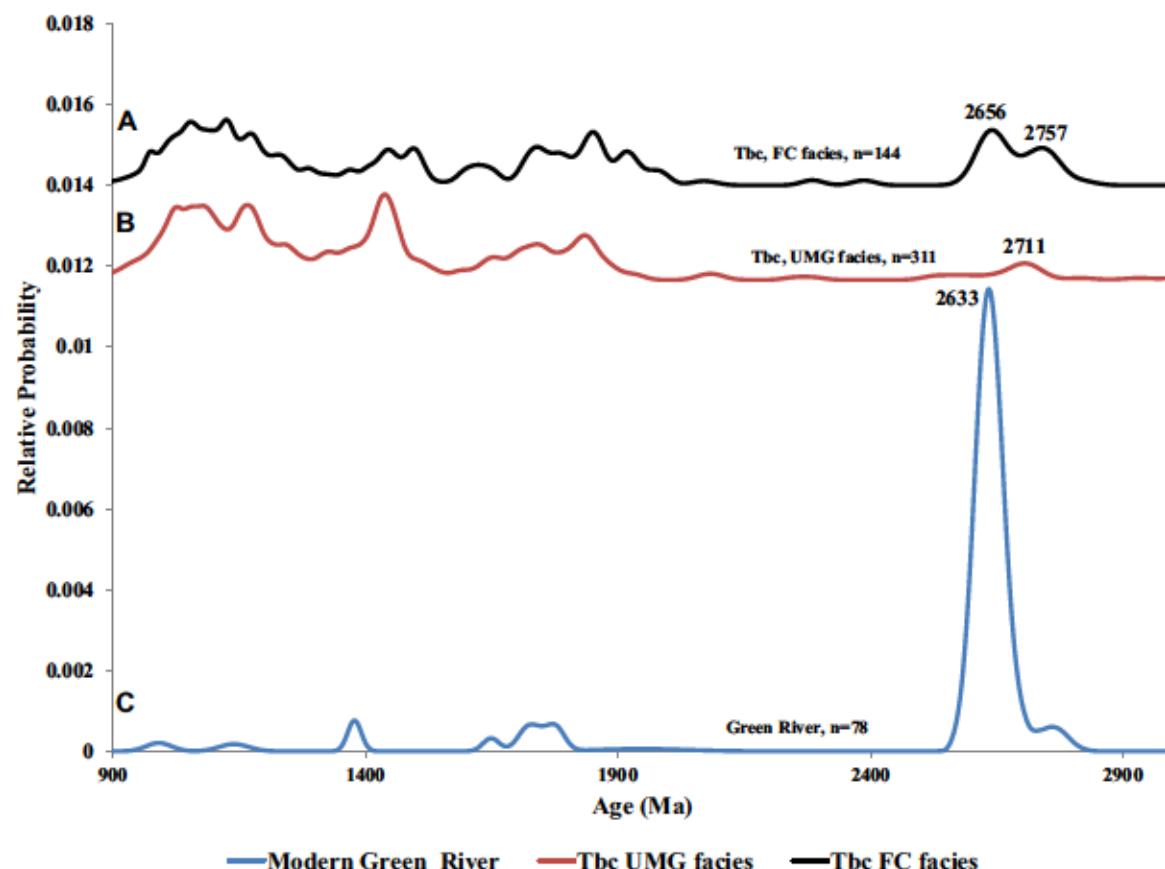


Figure 7. Normalized probability curves comparing detrital zircon age distributions for Archean and Proterozoic (2900–900 Ma) grains for the FC (Firehole Canyon) and UMG (Uinta Mountain Group) facies. (A, B) Bishop Conglomerate (Tbc). (C) Modern Green River. The FC and UMG curves are composites of the same samples as in Figure 6. Sample locations are shown in Figures 3B and 4.

Maximum Depositional Ages

Bishop Conglomerate

Detrital zircon and sanidine MDAs of the Bishop Conglomerate confirm that this unit is late Eocene–early Oligocene (Fig. 9; Table 1). In the southern Green River Basin, the zircon U–Pb MDA for the UMG facies is 34.7 ± 4.7 Ma (Firehole Canyon; Fig. 9), and the age range for the FC facies is 35.5 ± 0.6 – 33.4 ± 0.3 Ma (GRAT, Firehole Canyon; Fig. 9). The MDA estimates for the UMG and FC facies overlap, consistent with the observed stratigraphic interfingering of these two units.

In the Browns Park and western Sand Wash Basin regions, the zircon U–Pb MDAs for the UMG facies range from 32.2 ± 4.4 to 27.1 ± 1.2 Ma (Powder Wash, Sand Wash; Fig. 9). The 27.1 ± 1.2 Ma age at Sand Wash overlaps with the

detrital zircon U–Pb MDA of 27.6 ± 0.7 Ma calculated by Fan et al. (2015) for the Bishop Conglomerate outcrop along the nearby Little Snake River (Figs. 10A, 10B).

The zircon U–Pb MDA for the sandy facies of the Bishop Conglomerate at Antelope Butte (31.2 ± 0.6 Ma) is consistent with its stratigraphic position above the other two facies, which have older MDAs (Fig. 9). The zircon U–Pb MDA in the eastern Uinta Mountains (upper Crouse Canyon) is 27.0 ± 0.3 Ma, which is broadly consistent with the 26.2 ± 0.7 Ma K–Ar age of a tuff located ~90 m above the base of the Bishop Conglomerate on the Diamond Plateau near Vernal, Utah (Damon, 1970) (Fig. 9). In summary, the Bishop Conglomerate detrital zircon U–Pb-based age estimates (ca. 36–27 Ma) are remarkably consistent with the $^{40}\text{Ar}/^{39}\text{Ar}$ (ca. 34–30 Ma) and K–Ar (29–26 Ma) age estimates of Bishop Conglomerate tuffs from the southern flank of the Uinta Mountains (Damon, 1970; Hansen, 1986; Kowallis et al., 2005).

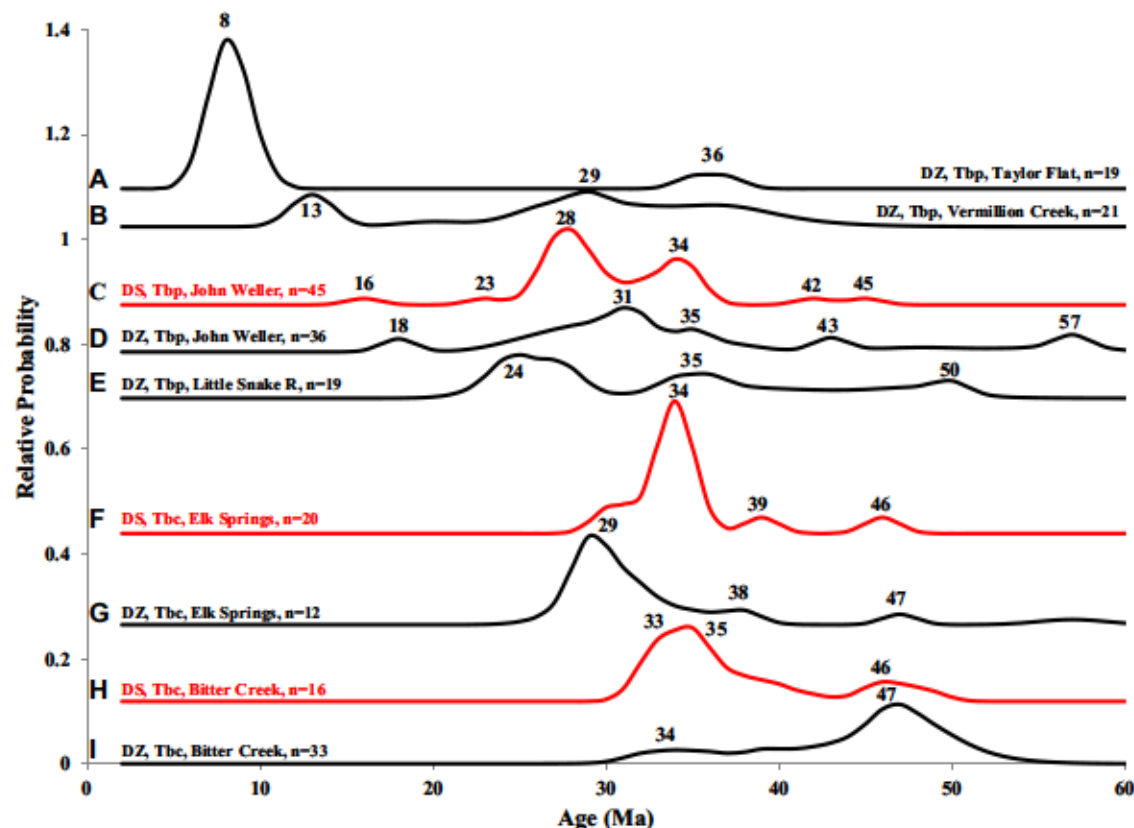


Figure 8. Normalized probability curves comparing detrital zircon (black curves) and sanidine (red curves) age spectra for young (0–60 Ma) grains for selected Browns Park Formation and Bishop Conglomerate samples. Sample locations are shown in Figure 3B. Paired samples C–D, F–G, and H–I represent detrital zircon U–Pb age and detrital sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ age data from the same sample. Numbers in bold are ages (in Ma) of major U–Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ age distributions; n—number of grains used in the probability plot. DZ—detrital zircon, DS—detrital sanidine, Tbp—Browns Park Formation, Tbc—Bishop Conglomerate.

Browns Park Formation

Detrital zircon and detrital sanidine MDAs of the Browns Park Formation confirm that the unit is late Oligocene to Miocene (ca. 25–8 Ma) (Izett, 1975; Luft, 1985), and suggest that the youngest sediments of the graben fill are concentrated in westernmost Browns Park along the Colorado-Utah border (Fig. 9; Table 1). The oldest zircon U–Pb MDA (24.5 ± 1.1 Ma) is from a tuff near the base of the Browns Park Formation, which is located ~20 m above the Bishop Conglomerate (MDA = 27.6 ± 0.7 Ma; Fan et al., 2015) (Fig. 10B). This tuff was dated previously to 24.8 ± 0.8 Ma using K–Ar dating methods (Izett, 1975). A zircon U–Pb MDA of 18.5 ± 1.8 Ma was obtained on pebbly tuffaceous sandstone directly overlying probable Bishop Conglomerate near Lodore Canyon. A bedded vitric tuff near Vermillion Creek produced an MDA of 13.7 ± 2.9 Ma. This tuff was dated previously to 9.1 ± 1.0 Ma (Izett, 1975) and 9.9 ± 0.4 Ma (Naeser et al., 1980) using fission track methods. Because thermal heating can anneal tracks, the fission track data probably represent a minimum depositional age.

The youngest MDAs of Browns Park Formation samples are from tuffaceous siltstone and vitric tuff located west of Vermillion Creek (Fig. 9; Table 1). Other than the ca. 14 Ma Vermillion Creek tuff, the oldest age estimate (9.2 ± 1.4 Ma) from this area comes from the lowermost of the two prominent vitric tuffs exposed at Jesse Ewing Canyon. This tuff is probably the one reported in Winkler (1970), and is dated to 11.8 ± 0.4 Ma using volcanic glass and K–Ar geochronology. The other two samples (Crouse Canyon, Taylor Flat) represent age estimates that are from stratigraphic positions that are younger than the Jesse Ewing Canyon tuff. The Taylor Flat sample is from a small remnant of the Browns Park Formation (elevation 1865 m) that is preserved beneath erosionally resistant river gravel of the ancestral Green River (Counts, 2005), and the sample is from a tuffaceous siltstone that overlies the Neoproterozoic Uinta Mountain Group. This remnant is significantly higher in elevation than any nearby Browns Park Formation outcrop, and probably represents the best estimate for the minimum age (8.4 ± 0.3 Ma) of the Browns Park Formation in the region.

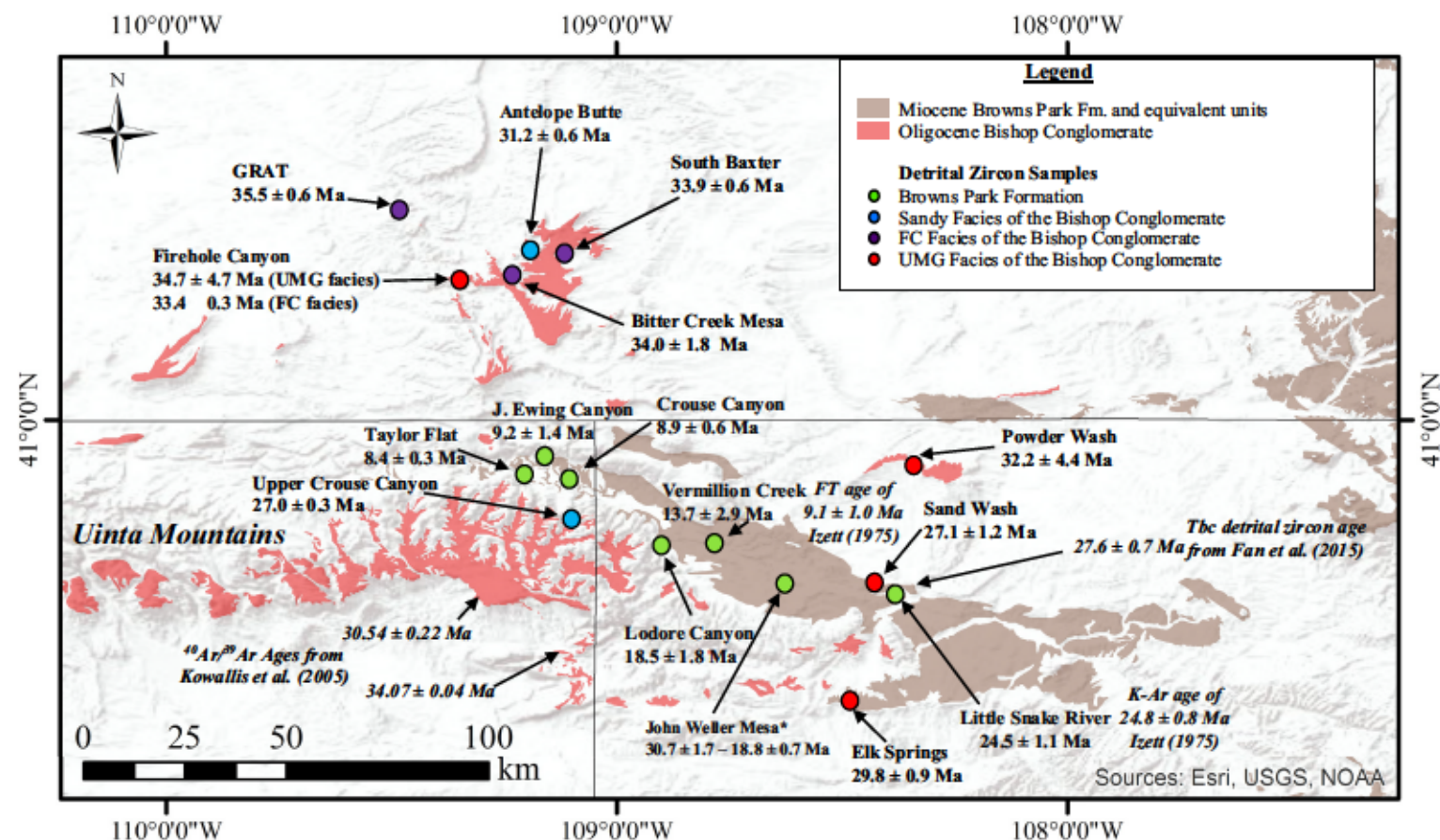


Figure 9. Outcrop map and summary of Bishop Conglomerate and Browns Park Formation (Fm.) maximum depositional age (MDA) data. Data are also presented in Table 1. Radiometric age estimates of the Bishop Conglomerate from Kowallis et al. (2005) and Fan et al. (2015) and of the Browns Park Formation from Izett (1975) are also shown. FT—fission track. Note that the MDA shown for John Weller Mesa (30.7 ± 1.7 Ma) was calculated using the youngest group of zircon U-Pb ages, but the youngest grain (18.8 ± 0.7 Ma), which was an outlier in this set of young grains, is also shown. Legend and detrital zircon sample abbreviations are as in Figure 3B. Sources: Esri—<https://www.esri.com/>; USGS—U.S. Geological Survey; NOAA—U.S. National Oceanic and Atmospheric Administration.

Comparison of Detrital Zircon and Sanidine Data

For the Elk Springs and Bitter Creek Bishop Conglomerate samples (Figs. 8F–8I), the young (younger than 60 Ma) detrital zircon and sanidine age distributions are similar with peaks of 35–33 Ma, 39–38 Ma, and 47–46 Ma. The main difference between the zircon U-Pb and sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ MDAs for the Bishop Conglomerate is the precision of the age estimates (Table 1). For example, the detrital sanidine MDA of the Bitter Creek sample is 33.4 ± 0.16 Ma, whereas the detrital zircon MDA of the same sample has a precision of ±1.8 m.y.

The young (younger than 60 Ma) detrital zircon and sanidine ages from the Browns Park Formation sample (John Weller Mesa) also show similarities, and age distributions cluster between 35 and 28 Ma (Figs. 8C, 8D). Detrital sanidine data show a younger (16 Ma) age distribution compared to the detrital zircon data (18 Ma). Both the zircon U-Pb and sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ MDAs (zircon = 30.7 ± 1.7 Ma; sanidine = 27.5 ± 0.3 Ma) for the John Weller Mesa are older than the youngest detrital grains (zircon = 18.8 ± 0.7 Ma; sanidine = 16.6 ± 0.05 Ma) in the sample (Table 1). The John Weller Mesa sample is located in the upper portion of the preserved Browns Park Formation section, and represents a sig-

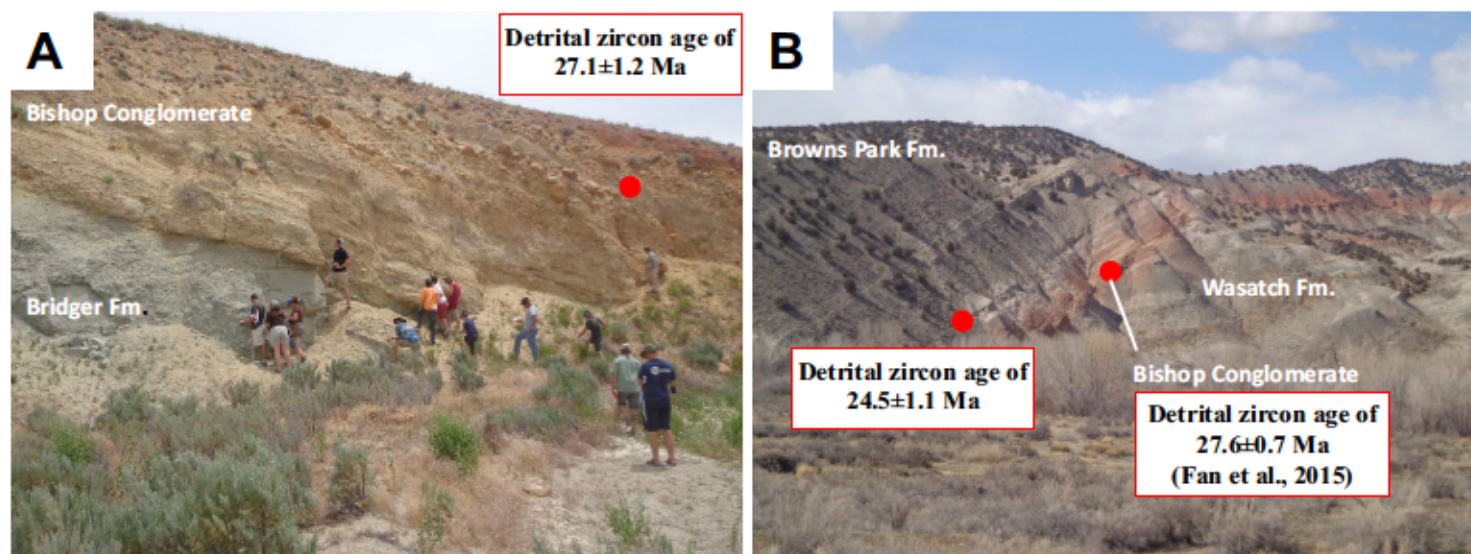


Figure 10. Field photographs of important stratigraphic and structural relationships in the Browns Park region of northwestern Colorado. (A) Sand Wash angular unconformity, looking southeast. Gray mudrock of the Bridger Formation (Fm.) dips north-northeast and is unconformably overlain by basal conglomerate and sandstone of the Bishop Conglomerate, which dips in the opposite direction (southwest) toward the axis of the Browns Park graben. Paleocurrent measurements from imbricated sandstone clasts in the Bishop Conglomerate show northeast-directed paleoflow, away from the eastern Uinta Mountains. The Uinta Mountains were the source of the light reddish-brown sandstone and reddish sandstone clasts that dominate the Bishop Conglomerate at this outcrop. The red dot is located near the detrital zircon sample site for Sand Wash [Bishop Conglomerate UMG (Uinta Mountain Group) facies]. See Figure 3B for location; note people for scale. (B) Little Snake River angular unconformity, looking northwest. Where the Little Snake River enters Browns Park, beds of the light gray Browns Park Formation and the underlying reddish Bishop Conglomerate dip southwest toward the axis of the Browns Park graben, and unconformably overlie more gently dipping strata of the Eocene Wasatch Formation. At this location, the Bishop Conglomerate is only 2–3 m thick and consists of a pebbly sandstone with abundant red sandstone clasts. Fan et al. (2015) reported an age of 27.6 ± 0.7 Ma from this locality as the basal age of the Browns Park Formation, but this red conglomeratic unit underlying the tuffaceous Browns Park Formation along the Little Snake River is generally regarded as part of the Bishop Conglomerate (Hansen, 1986). The younger Little Snake River detrital zircon sample was acquired from a light gray vitric tuff near the base of the Browns Park Formation. A K-Ar age of 24.8 ± 0.8 Ma for this tuff was reported by Izzett (1975).

nificantly higher position stratigraphically than nearby Bishop Conglomerate outcrops, so it is likely that the MDAs for this sample are older than its depositional age.

DISCUSSION

Volcanic Ash-Fall Origin for Younger than 40 Ma Detrital Zircon and Sanidine Grains

The abundance of younger than 40 Ma detrital zircon and sanidine grains in the Bishop Conglomerate and Browns Park Formation samples (Fig. 8) cannot be explained simply by traditional explanations for detrital grain spectra involving bedrock erosion and fluvial transport. For example, during the time frame (ca. 36–27 Ma) represented by the Bishop Conglomerate, there were no nearby late Eocene–Oligocene outcrops to explain the abundant younger than 40 Ma detrital zircon and sanidine grains (Love and Christiansen, 1985). The

youngest bedrock unit in the region was probably the Eocene Bridger Formation, which has a minimum age of ca. 47 Ma (Murphey and Evanoff, 2007). It is also unlikely that grains of this young (younger than 40 Ma) age were transported by rivers draining hinterlands composed of Oligocene and younger rocks, and deposited subsequently in southwestern Wyoming and northwestern Colorado. Rocks of this age are abundant in the southern Basin and Range, but barriers such as the Wyoming fold and thrust belt (Fig. 1) make a fluvial connection between the regions unlikely. Instead, the younger than 40 Ma detrital zircon and sanidine grains in the Bishop Conglomerate and Browns Park Formation probably represent a combination of Oligocene through Miocene volcanism, ash-fall deposition, and subsequent fluvial reworking. This interpretation is supported by a combination of stratigraphic observations and age estimates of Cenozoic volcanic activity in the western U.S.

Figure 8 (samples A, B, D, E, G) shows that for a composite stratigraphic succession of the Bishop Conglomerate and Browns Park Formation detrital zircon samples, the age of the youngest U-Pb age distributions decrease systematically at successively younger stratigraphic levels. This pattern is consis-

tent with episodic Bishop Conglomerate and Browns Park Formation deposition accompanied by periodic ash-fall deposition and fluvial reworking. This interpretation is further supported by comparing the timing of Cenozoic ignimbrite events with the general ages of the young detrital grains.

Figure 11 compares the timing and volume of Oligocene and Miocene ignimbrites (Perkins et al., 1998; Cather et al., 2009; Best et al., 2013; Lipman and Bachmann, 2015) with detrital zircon data for the Browns Park Formation and the Bishop Conglomerate. There is a good correspondence between the timing of the Cenozoic ignimbrite flux for western North America and the ages of the dominant detrital zircon age distributions in the Bishop Conglomerate, and to a lesser extent, in the Browns Park Formation (Fig. 11A). An even better match between the ignimbrite eruptive flux curve and the Browns Park For-

mation detrital zircon age distributions is expected if the flux curve were to incorporate data for the Snake River Plain volcanic field, which ranges from ca. 16 to 6 Ma (Perkins et al., 1998; Perkins and Nash, 2002). Based on proximity to the study area, it is likely that the younger than 20 Ma detrital zircon and sanidine grains in the Browns Park Formation (Fig. 8, samples A–D) were derived originally as ash fall from this volcanic field.

Previous studies have suggested that Basin and Range volcanism was a primary source for both the Browns Park Formation (Luft, 1985) and Bishop Conglomerate (Kowallis et al., 2005) tephra. For example, geochemical studies and radiometric dating of a distal tuff bed in the Bishop Conglomerate near Vernal, Utah, suggested that the Cottonwood Wash ignimbrite (age 31.13 Ma, estimated volume of 2000 km³; Fig. 11B) was the source of the tuff (Kowallis

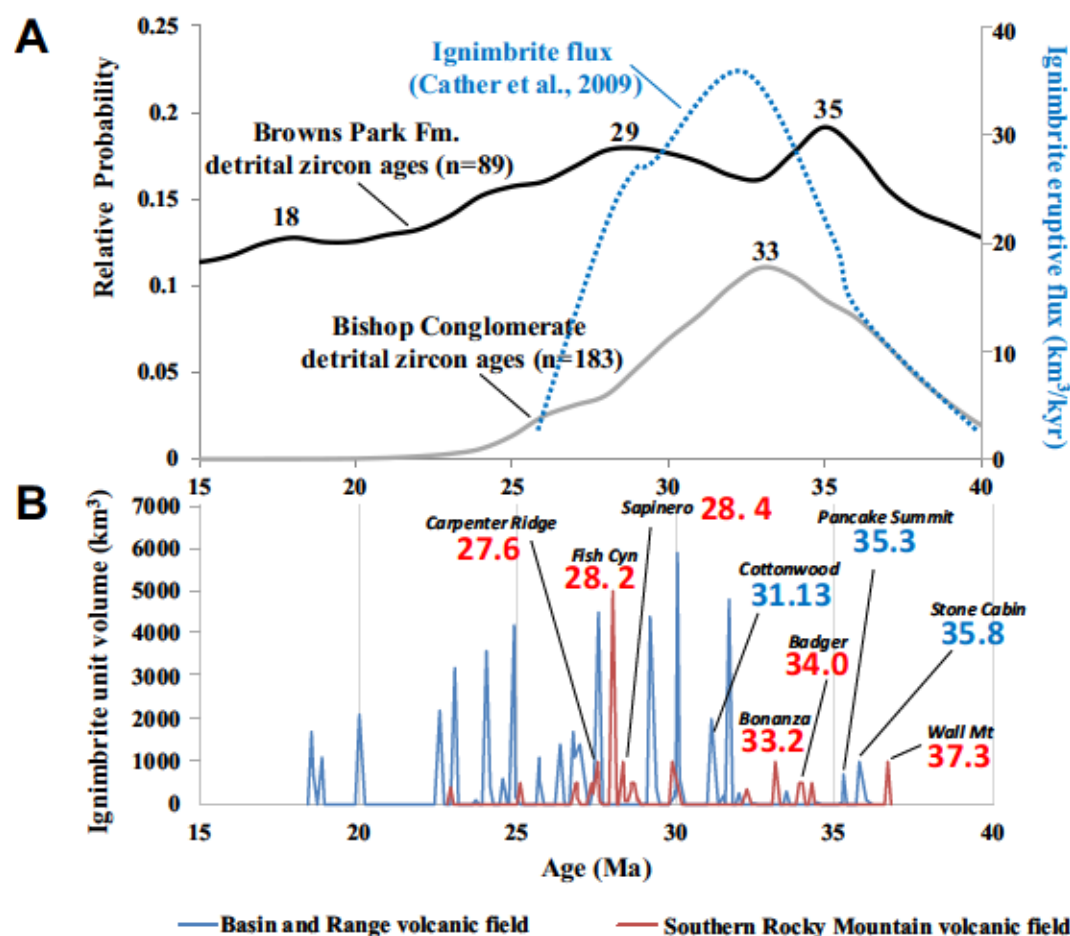


Figure 11. (A) Normalized probability curves of young (younger than 40 Ma) detrital zircon U-Pb ages for all Browns Park Formation samples (black curve) and Bishop Conglomerate samples (gray curve; n is number of zircon grains used to construct the curves). These curves are compared to the Cenozoic eruptive flux curve for western North America (Cather et al., 2009). (B) Summary of timing of Oligocene-Miocene ignimbrite eruptions of the Basin and Range (blue curve) and Southern Rocky Mountain (red curve) volcanic fields. Data are from Perkins et al. (1998), Best et al. (2013), and Lipman and Bachmann (2015). Radiometric ages of important ignimbrites are shown as blue numbers for the Basin and Range volcanic field and red numbers for the Southern Rocky Mountain volcanic field.

et al., 2005). Comparison of the detrital zircon U-Pb age distributions with the timing of Basin and Range and San Juan Mountain ignimbrite events suggests only a general correspondence (Fig. 11B). However, the precision of the detrital sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology provides a better means for correlating Bishop Conglomerate and Browns Park Formation sanidine grains with specific caldera-forming eruptions (Fig. 12). Detrital sanidine ages and K/Ca data for the Bishop Conglomerate Bitter Creek sample match well with the age and K/Ca data for tuffs for the Stone Cabin Tuff in the Basin and Range Central Nevada ignimbrite field (CNIF) (Best et al., 2013), and somewhat well with the age and K/Ca data for the Basin and Range Pancake Summit Tuff of the CNIF (Best et al., 2013) (Fig. 12A). The K/Ca data for the Wall Mountain Tuff of the Southern Rocky Mountain volcanic field (SRMVF), however, do not match well with the Bishop Conglomerate detrital sanidine data. The age and K/Ca data for the Browns Park Formation John Weller Mesa sample matches well with data for the Carpenter Ridge, Fish Canyon, Sapinero Mesa, and Badger Creek Tuffs of the SRMVF (Fig. 12B). These results confirm that Basin and Range volcanism contributed to the Bishop Conglomerate sedimentation (Kowallis et al., 2005). Contrary to previous interpretations (e.g., Luft, 1985), however, the

detrital sanidine data suggest that the SRMVF, along with the Snake River Plain and Basin and Range volcanic fields, was a source for a portion of the Browns Park Formation tephra.

Prior studies have proposed that Oligocene–Miocene volcanism played a significant role in the accumulation of fine-grained Cenozoic units in the western U.S. based on a combination of radiometric dating, analysis of volcanic glass chemistry, and detrital zircon studies (Blaylock, 1998; Larson and Evanoff, 1998; Fan et al., 2015; Rowley and Fan, 2016). Larson and Evanoff (1998) used radiometric ages to broadly correlate Oligocene White River Formation tuffs located in eastern Wyoming and western Nebraska with Basin and Range volcanism. Blaylock (1998) refined tephra correlations of Larson and Evanoff (1998) using detailed geochemical analyses of comagmatic minerals, but also concluded that the southern Basin and Range was the source of several White River Formation tephra units. Fan et al. (2015) used overall similarities in the timing of Oligocene volcanism and MDAs of detrital zircon samples to explain the abundance of Oligocene detrital zircon grains in tuffaceous fluvial deposits in Wyoming and Nebraska. This study suggested, however, that the Sierra Madre Occidental volcanic field in Mexico may have been

A Bishop Conglomerate, Bitter Ck B Browns Park Fm., John Weller Mesa

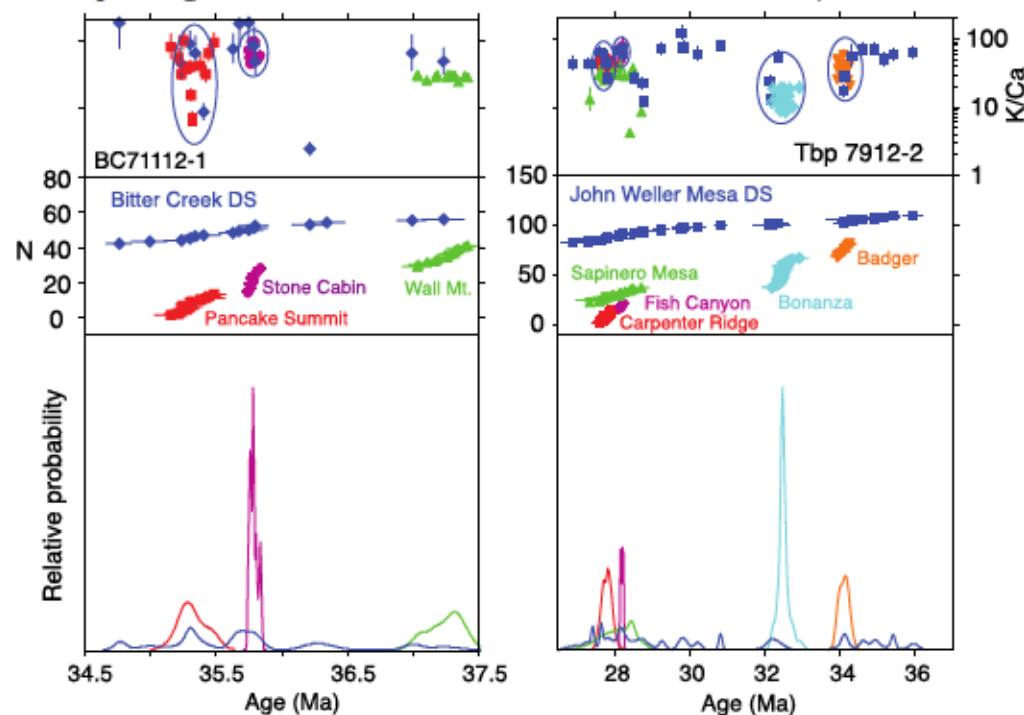


Figure 12. Detrital sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ and tuff geochronology. (A) Detrital sanidine (DS) data from the 3 m.y. interval 37.5–34.5 Ma for the Bishop Conglomerate (Bitter Creek (BC) sample, location shown in Fig. 3B) compared to possible caldera sources from the Basin and Range (Stone Cabin, Pancake Summit) and Southern Rocky Mountain (Wall Mountain) ignimbrite units. Bishop Conglomerate detrital sanidine ages and K/Ca values match well with data for the Stone Cabin and Pancake Summit Tuffs, but K/Ca values do not match those of the Wall Mountain Tuff. (B) Detrital sanidine data from 36 to 26 Ma for the Browns Park Formation (Tbp; John Weller Mesa sample; location shown in Fig. 3B) compared to data for Southern Rocky Mountain ignimbrite units. Browns Park Formation detrital sanidine ages and K/Ca values match well with data for Southern Rocky Mountain ignimbrite units including the Sapinero Mesa, Fish Canyon, Carpenter Ridge, and Badger Creek Tuffs. Ck—Creek; N—number of analyses.

the source for the Oligocene detrital zircon grains. Figure 13 summarizes our new interpretations of Oligocene–Miocene ash-fall deposition based on the data and discussion presented in this study. In summary, far-traveled tephra that mantled the landscapes of southwestern Wyoming, northeastern Utah, and northwestern Colorado during the latest Eocene through Miocene was reworked subsequently into Bishop and Browns Park depositional systems.

Repetition of this sequence of events produced deposits with successively younger detrital zircon grains, and further suggests that MDAs for the Bishop Conglomerate and Browns Park Formation are close to true depositional ages, as has been suggested for other Cenozoic units in the western U.S. that accumulated in regions of semicontinuous ignimbrite volcanism (Fan et al., 2015; Rowley and Fan, 2016).

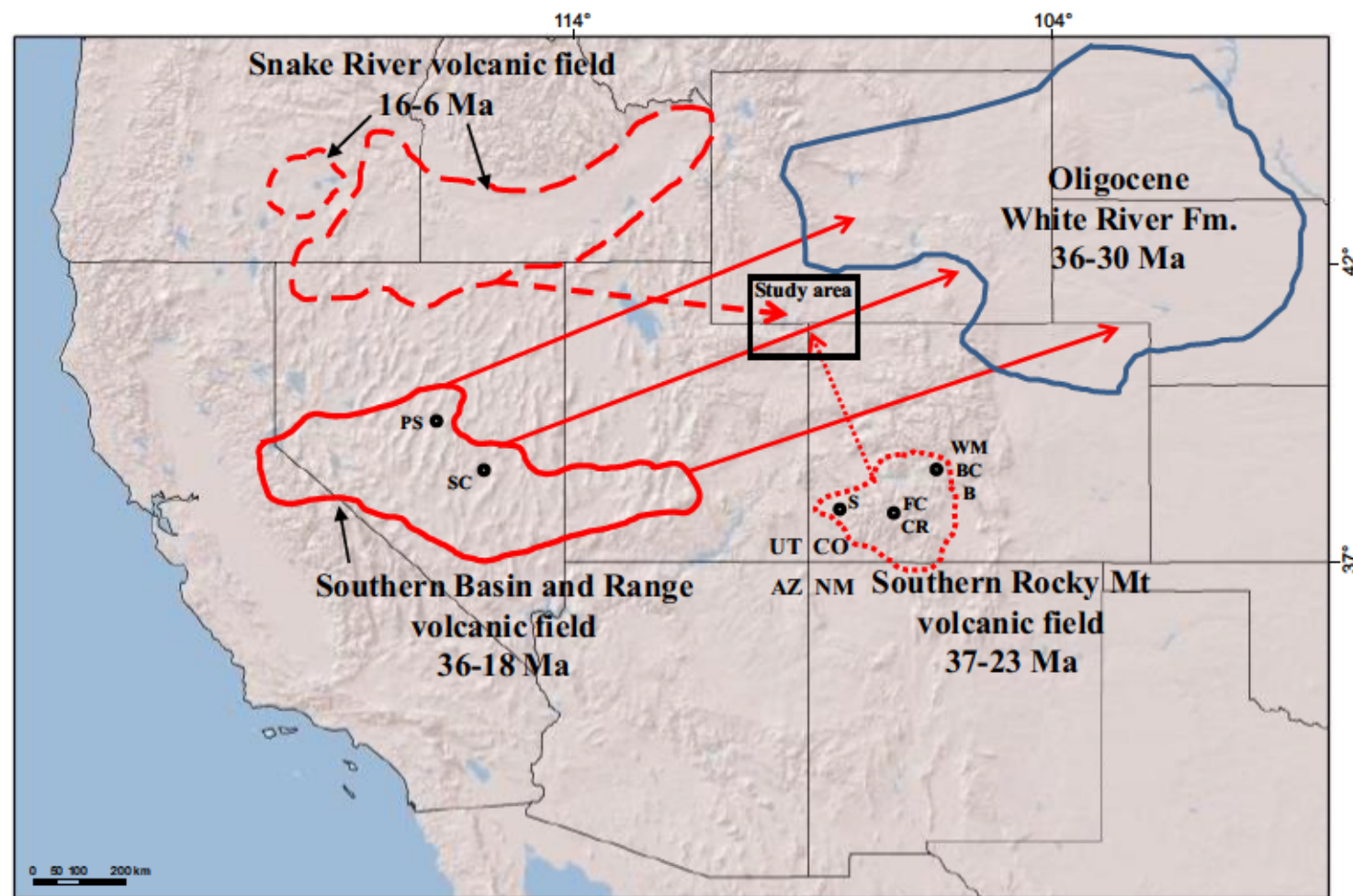


Figure 13. Map showing Oligocene–Miocene volcanic centers including the Snake River, southern Basin and Range, and Southern Rocky Mountain volcanic fields (solid and dashed red lines), paleowind directions (solid and dashed red arrows), the approximate distribution of tuffaceous Oligocene White River Formation deposits (blue solid line), and the study area (black box). Black circles mark the approximate locations of calderas associated with ignimbrite units referred to in the text. Abbreviations refer to ignimbrite units associated with the calderas: B—Bonanza, BC—Badger Creek, CR—Carpenter Ridge, FC—Fish Canyon, PS—Pancake Summit, S—Sapinero Mesa, SC—Stone Cabin, and WM—Wall Mountain Tuffs. Data were compiled from Best et al. (1989, 2013), Larson and Evanoff (1998), Swisher and Prothero (1990), Obradovich et al. (1995), Perkins et al. (1998), Perkins and Nash (2002), and Lipman and Bachmann (2015). AZ—Arizona; CO—Colorado; NM—New Mexico; UT—Utah.

Provenance and Paleogeography during the Paleogene

Differences in the distribution, paleocurrents, and composition of the Bishop Conglomerate UMG and FC facies suggest that these deposits represent tributaries and a mainstem river, respectively, of a latest Eocene to early Oligocene fluvial system. Previous investigators have shown that the Bishop Conglomerate (UMG facies) represents a series of rivers that drained radially away from the Uinta Mountains (Bradley, 1936; Hansen, 1986). The abundance of reddish sandstone clasts and Grenville-age grains in the UMG facies confirms that the provenance was the Neoproterozoic UMG in the nearby Uinta Mountains. New numerical age constraints show that this river system flowed along the north flank of the Uinta Mountains ca. 36–27 Ma (Table 1).

Paleocurrent data and topographic relationships indicate that the FC facies represents a distinctly different river system from the UMG facies rivers. At Firehole Canyon, the UMG and FC facies interfinger, and paleocurrent directions for the two facies are roughly perpendicular to one another (Fig. 4), which suggests that this area was a paleoconfluence. At South Baxter, the FC facies outcrop belt occupies a distinctive topographic saddle that is bordered to the south by Miller Mountain, which represents northward-sloping Bishop Conglomerate UMG facies, and to the north by Aspen Mountain, which represents southward-sloping Bishop Conglomerate sandstone-dominated gravels (Fig. 4). The north-south orientation of the saddle suggests that the FC facies was part of an east-west-trending segment of a lower order, probable mainstem river, which was fed by Bishop tributaries. The basal elevation of the FC facies decreases ~90 m to the west over a distance of ~20 km between South Baxter and Firehole Canyon. This observation could indicate that the mainstem river flowed to the west, but this interpretation assumes that there has been no significant post-Oligocene deformation of the FC facies conglomerates.

The presence of volcanic clasts and the abundance of rounded quartzite cobbles in the FC facies clearly show that the provenance of this facies differs from the UMG facies. Figure 14 summarizes competing possibilities for the source areas and flow directions of the Bishop river system representing these facies. Possible sources of the quartzite clasts in the FC facies include Proterozoic quartzite in the Sierra Madre Mountains (Scott et al., 2011) as well as Cretaceous and Paleocene quartzite conglomerates of the Wyoming fold and thrust belt (DeCelles, 1994) and northwestern Wyoming (Lyndsey, 1972). Intermediate volcanic clasts in the FC facies also have several potential sources. These include upper Cretaceous and Paleocene quartzite conglomerates located near Jackson Hole, Wyoming, that contain a significant proportion (as much as 15%) of intermediate volcanic clasts (Lyndsey, 1972), and the Eocene Challis and/or Absaroka volcanic fields of Idaho and Wyoming, respectively. Previous studies suggest that a river drained the Challis and/or Absaroka volcanic fields and flowed into southwestern Wyoming in the Eocene (ca. 50–47 Ma) (Carroll et al., 2008; Davis et al., 2008; Chetel et al., 2011). It is possible that a segment of the Bishop river system similarly drained the Challis and/or Absaroka volcanic fields, or perhaps it reworked the older volcanic and quartzite conglomerates of northwestern Wyoming. Alternatively, Roehler (1973, 1992) noted that Eo-

cene Washakie Formation conglomeratic sandstones (Adobe Town Member) that crop out in the Washakie Basin to the east of the FC facies outcrop belt have minor quantities of intermediate volcanic clasts. It is conceivable that a west-flowing Bishop Conglomerate river reworked conglomerate of the Washakie Formation to produce the minor volcanic component observed in the FC facies (Fig. 14).

The absence of granitic clasts in the FC facies conglomerate is important because granitic clasts would be expected for a river draining either the Wind River or Sierra Madre Mountains (Fig. 14). Detrital zircon data provide a possible link between the provenance of the FC facies and the Wind River Mountains. The FC facies contains a greater abundance of ca. 2650 Ma detrital zircon grains than the UMG facies, and Archean detrital zircon grains are similarly abundant in the modern Green River, which drains the Wind River Mountains (Fig. 7). Alternatively, the absence of granitic clasts in the FC facies could indicate that basement rocks were buried by younger sediment during the time represented by the Bishop rivers (Fan et al., 2015; Rowley and Fan, 2016).

In summary, the Bishop Conglomerate represents the most significant late Eocene–Oligocene river system of the Uinta Mountains region, but the interpretation of the flow direction of this river is unresolved. An east- or south-east-flowing late Eocene–early Oligocene river, as envisioned by Hansen (1986) and Fan et al. (2015), is a possible interpretation for the Bishop Conglomerate FC facies, but it would require post-Oligocene, down-to-the-west tilting to explain the present-day elevations of the conglomerate, and it conflicts with FC facies paleocurrent data. A west-flowing river is consistent with FC facies paleocurrents and conglomerate elevations, but the absence of granitic clasts from the Sierra Madre Mountains is problematic. Lucchitta et al. (2011) and Ferguson (2011), as summarized in Cather et al. (2012), argued for possible north-flowing rivers that extended across the Colorado Plateau and possibly into Wyoming in the vicinity of Powder Wash (Fig. 14). However, the suggested ages (late Oligocene to Pliocene) for these rivers are significantly younger than those proposed for the Bishop river system, and deposits of north-flowing rivers draining Laramide uplifts of the Colorado Rocky Mountains should contain granitic clasts. For these reasons, it is unlikely that the Bishop river system is part of the north-flowing river envisioned by Lucchitta et al. (2011) and Ferguson (2011).

Timing of Collapse of the Eastern Uinta Mountains

Detrital zircon U–Pb ages of the Bishop Conglomerate and Browns Park Formation samples provide new constraints on the timing of the collapse of the eastern Uinta Mountains (Hansen, 1969a, 1984, 1986). Based on stratigraphic and structural relationships in the Browns Park region, Hansen (1986) suggested that the main phase of collapse of the eastern Uinta Mountains occurred between deposition of the Oligocene Bishop Conglomerate and the Miocene Browns Park Formation. However, tilting and faulting of the Browns Park Formation strata show that deformation continued through at least Miocene time. Late Oligocene to early Miocene collapse is based on three key

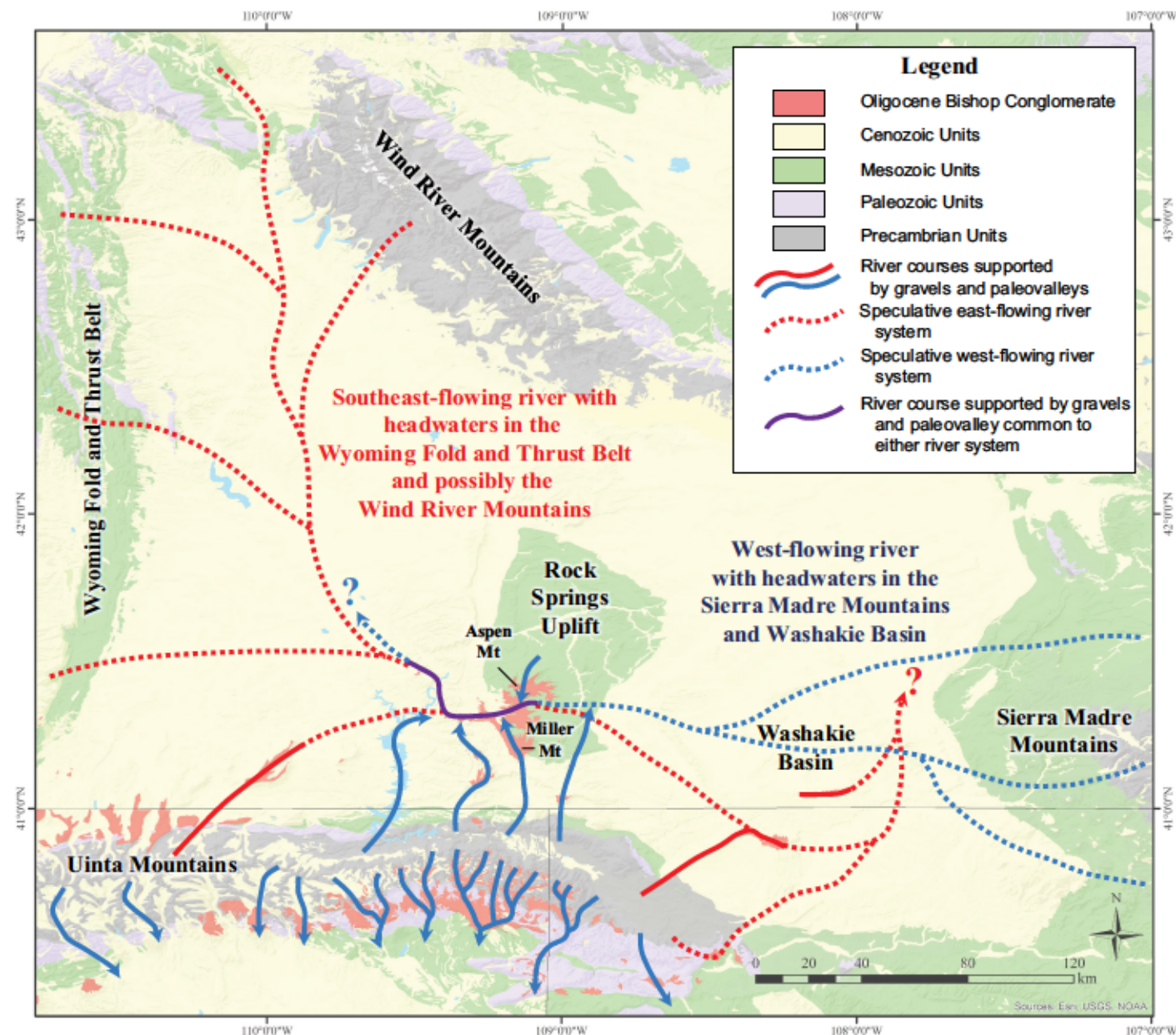


Figure 14. Map showing possible courses of the late Eocene to early Oligocene (ca. 36–27 Ma) Bishop river system in the southern Green River Basin of Wyoming and adjacent areas. The river system could have flowed southeast across southwestern Wyoming with headwaters in the Wyoming fold and thrust belt and Wind River Mountains (red lines). Alternatively, the river system could have flowed west with sources in the Sierra Madre Mountains (blue lines).

observations. First, the Bishop Conglomerate deposits extend radially away from the structural axis of the eastern Uinta Mountains, including areas beyond the present-day Browns Park graben (Fig. 15). This observation clearly demonstrates that the graben did not exist at the time of Bishop Conglomerate deposition. Second, a spectacular series of barbed drainages present along the south flank of the eastern Uinta Mountains, especially near Diamond Plateau, provide additional evidence of post-Bishop Conglomerate collapse

(Hansen, 1984, 1986). The barbed drainages consist of southward-draining Bishop Conglomerate paleovalleys reincised by modern streams. The modern streams drain southward from their headwaters, but then turn sharply to the north where they join the Green River in present-day Browns Park. Hansen (1986) pointed out that this reversal in drainage direction clearly postdated the Bishop Conglomerate, and suggested that collapse of Oligocene topography, on the order of several hundred meters (Bradley, 1936; Hansen, 1986), initiated

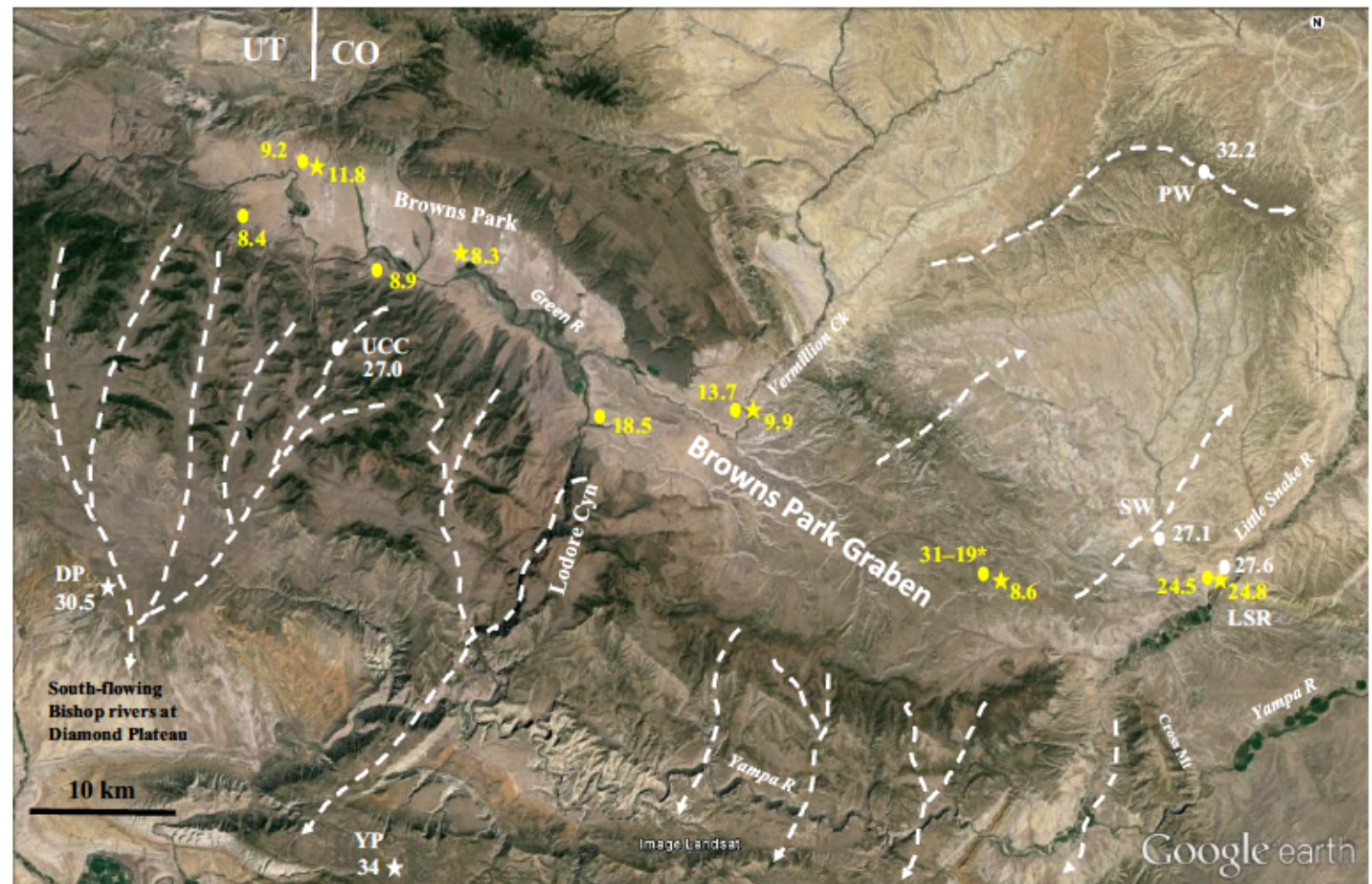


Figure 15. Satellite image of the Browns Park area in northeastern Utah (UT) and northwestern Colorado (CO) showing possible Oligocene Bishop Conglomerate river courses (white dashed lines with arrows pointing in the direction of paleoflow). White numbers are age estimates (in Ma) for the Bishop Conglomerate based on detrital zircon U-Pb maximum depositional ages (white dots) and radiometric ages (white stars). Radiometric $^{40}\text{Ar}/^{39}\text{Ar}$ age estimates are from Kowallis et al. (2005). Yellow numbers are age estimates (in Ma) for the Browns Park Formation (Fm.) based on detrital zircon U-Pb maximum depositional ages (yellow dots) and a combination of fission track and K-Ar age estimates (yellow stars). K-Ar and fission track data are from Izett (1975) and Luft (1985). DP—Diamond Mountain Plateau, YP—Yampa Plateau, UCC—upper Crouse Canyon, SW—Sand Wash, LSR—Little Snake River, PW—Powder Wash. Image from Google Earth.

the formation of the Browns Park graben. Third, the transition from Bishop Conglomerate to Browns Park sedimentation accompanied the collapse of the eastern Uinta Mountains (Hansen, 1986). In this scenario, late Oligocene to early Miocene Browns Park sediments began to accumulate in Browns Park due to a combination of changes in topography, disruption of the Bishop Conglomerate drainage networks, and the formation of the Browns Park graben, which provided accommodation space for Browns Park sediments. Browns Park Formation deposits are similarly concentrated in grabens elsewhere in northwestern Colorado and along the Colorado-Wyoming border (Izett, 1975; Luft, 1985; Buffler, 2003).

Browns Park Formation facies and age relationships in the Browns Park graben further support the idea that the transition from coarse-grained Bishop Conglomerate fluvial deposition to fine-grained, tuffaceous Browns Park sedimentation is consistent with extensional collapse in the late Oligocene. In the vicinity of the Little Snake River, excellent exposures of the lower Browns Park Formation show significant numbers of siliceous, micritic, and sometimes oolitic limestone beds within the lowermost 100 m of the formation. The limestones along with the presence of broad lenses of reworked tuff suggest that shallow ponds existed during the time represented by lower Browns Park sedimentation. Thick, lenticular fluvial sand bodies are conspicuously absent in these laterally continuous exposures. Collectively, these observations support the interpretation of internal drainage within a closed basin during the early stages of Browns Park sedimentation in response to the formation of the Browns Park graben. Alternatively, the transition from Bishop Conglomerate to Browns Park sedimentation could reflect changes in climate, and concomitant changes in sediment composition and caliber. However, the age constraints presented in this study suggest that the transition from Bishop Conglomerate fluvial conglomeratic to Browns Park fluvial-lacustrine deposition occurred at different times in different places, as discussed in the following section, and changes due to climate are likely to be synchronous across a region.

Age Constraints

New detrital zircon MDAs constrain the timing of the structural and geomorphic changes to after the late Oligocene. Exposures of the lower Browns Park Formation along the northern flank of the Little Snake River valley (Fig. 10B) include a vitric tuff located ~20 m above the base of the Bishop Conglomerate (LSR in Fig. 15). The detrital zircon MDA of 24.5 ± 1.1 Ma supports a previous K-Ar radiometric age of 24.8 ± 0.8 Ma (Izett, 1975) from approximately the same location. These data indicate that lower Browns Park Formation sediments began to accumulate by ca. 25 Ma, and based on the reasoning outlined previously, ca. 25 Ma represents the minimum age for collapse of the eastern Uinta Mountains. Subsequent deformation has tilted both the Bishop Conglomerate and Browns Park Formation beds to the southwest (Fig. 10B).

The youngest Bishop Conglomerate samples provide additional constraints on the timing of structural collapse in the region. At Powder Wash (PW

in Fig. 15), the presence of Bishop Conglomerate northeast of the Browns Park region shows that this stream system predates collapse of the eastern Uinta Mountains. A new 32.2 ± 4.4 Ma MDA for these sediments provides a maximum age for the timing of collapse. Of even greater importance, however, is the new age estimate and structural relations at Sand Wash (Fig. 10A; SW in Fig. 15). At this location, Bishop Conglomerate is represented by imbricated sandy conglomerates and conglomeratic sandstones with a detrital zircon MDA of ca. 27.1 ± 1.2 Ma. This age is consistent with a similar detrital zircon MDA of 27.6 ± 0.7 Ma reported by Fan et al. (2015) for the Bishop Conglomerate along the nearby Little Snake River (Figs. 10B and 15). Imbricated clasts within the Bishop Conglomerate indicate northeast-directed paleoflow of the Bishop river system at Sand Wash. However, the Bishop Conglomerate beds at Sand Wash dip southwest, toward the structural axis of the Browns Park graben. Collectively, these observations suggest that collapse of the eastern Uinta Mountains and cessation of Bishop Conglomerate deposition occurred sometime after ca. 28 Ma.

An additional constraint on the timing of the collapse of the eastern Uinta Mountains is provided by a new age estimate of the Bishop Conglomerate sandy facies in upper Crouse Canyon (UCC in Fig. 15). In this area, a few tens of meters of friable, light gray tuffaceous sandstone and siltstone partially fill a Bishop Conglomerate paleovalley, which is part of the Diamond Plateau dendritic channel network showing southwest-directed paleoflow. The color, texture, and friable consistence of the deposits resemble those of the Browns Park Formation, but these sandy units are mapped as Bishop Conglomerate (Sprinkel, 2006). The detrital zircon MDA of the upper Crouse Canyon sandstone is 27.0 ± 0.3 Ma. This age estimate is older than the oldest Browns Park Formation age in the Browns Park graben, overlaps in age with conglomeratic facies of the Bishop Conglomerate along the Little Snake River and at Sand Wash, and is only slightly younger than a radiometric age estimate (ca. 30.5 Ma) for sandy, tuffaceous deposits overlying conglomerates of the Bishop Conglomerate on the Diamond Plateau (Kowallis et al., 2005; DP in Fig. 15). Kowallis et al. (2005) dated the main body of the Bishop Conglomerate to ca. 34 Ma (YP in Fig. 15). Collectively, these observations suggest that the transition from coarse-grained to fine-grained Bishop Conglomerate deposition occurred ca. 34–27 Ma, along the southeastern flank of the Uinta Mountains. Along the northeastern flank of the Uinta Mountains in southwestern Wyoming, this transition occurred at approximately the same time, ca. 34–31 Ma (Fig. 9). In contrast, the accumulation of sandy, tuffaceous sediment (Browns Park Formation) began ca. 28–25 Ma in Browns Park. The time transgressive nature of this lithologic transition supports the idea that collapse of the eastern Uinta Mountains beginning sometime after 28 Ma facilitated the accumulation of tuffaceous sediments in the Browns Park graben.

The distribution of Browns Park age estimates in Figure 15 shows an anomalous cluster of young (younger than 12 Ma) ages in westernmost Browns Park along the Green River upstream of Lodore Canyon. In this region, late Miocene Browns Park sediments unconformably onlap Neoproterozoic UMG sandstone beds; the Bishop Conglomerate is absent. The absence

of older (older than 12 Ma) Browns Park Formation and Bishop Conglomerate in this area raises the possibility that this region of westernmost Browns Park collapsed more recently than areas located southeast of Lodore Canyon and Vermillion Creek where older sediments are present (Fig. 15). One scenario is that collapse of the eastern Uinta Mountains occurred in a series of phases, the earliest (ca. 28–25 Ma) affecting areas between Vermillion Creek and the Little Snake River. Later in time, westernmost Browns Park collapsed, leading to the onset of a younger phase of Browns Park sedimentation ca. 12–8 Ma. This possibility could explain why it took the upper Green River until sometime between 8 and 2 Ma to integrate through this region (Aslan et al., 2013). Furthermore, the difference in the age estimates of the Browns Park Formation deposits in westernmost Browns Park (ca. 12–8 Ma) and lithologically similar Bishop Conglomerate sediments in upper Crouse Canyon (ca. 27 Ma) argues against the possibility that the Green River's course across the eastern Uinta Mountains through Lodore Canyon was due to stream capture of a superimposed river. If the Browns Park graben had been completely filled to the level of Bishop Conglomerate paleovalleys that flank the graben, then late Miocene Browns Park Formation rather than ca. 27 Ma sandy facies of the Bishop Conglomerate would fill the paleovalleys. Instead, the new age estimates lend support to the stream capture model postulated by Pederson and Haddor (2005) that invokes a combination of headward erosion and superimposition by the Green River.

Regional Comparisons

The importance of Neogene tectonism in modifying earlier uplifts is also seen in numerous areas. To the north of the Uinta Mountains, in central Wyoming, the southern margin of the Wind River Mountains is marked by east-southeast–west-northwest normal faults (e.g., Sales, 1983; Sutherland and Hausel, 2006). Precise age control for the timing of fault movement is inferred to be prekinematic to synkinematic to the deposition of the Miocene Split Rock Formation. To the east of this range, the former structural culmination of the Granite Mountains also foundered in north-northeast–south-southwest extension during deposition of the Miocene Moonstone Formation (see Sutherland and Hausel, 2003). The age of the Moonstone Formation is constrained by vertebrate paleontology, U–Pb zircon geochronology, and magnetostratigraphically (Prothero et al., 2008). In addition, there are east-southeast–west-northwest–striking normal faults on the Rock Springs uplift (Love and Christiansen, 1985) and Piceance Basin (Cashion, 1973) that currently lack age control, but may be associated with Miocene extension. For all of these locations and syn-tectonic deposits, regional uplift and drainage evolution may have been driven by post-Oligocene mantle convection in the Rocky Mountain region (Karlstrom et al., 2012; Rosenberg et al., 2014), delamination of lithospheric mantle beneath in western Utah (Levander et al., 2011) and central Colorado (Hansen et al., 2013), and a combination of regional plate-related(?) extension plus mantle forcings affecting the margins of the Colorado Plateau (Ricketts et al., 2016).

CONCLUSIONS

New sedimentological observations and a combination of detrital zircon and sanidine provenance data and age constraints provide insights on the paleogeographic and structural evolution of the Uinta Mountains region during the late Eocene to late Miocene (ca. 36–8 Ma). During the time represented by the Bishop Conglomerate (ca. 36–27 Ma), tributary streams drained radially away from the crest of the Uinta Mountains and, in the southern Green River Basin, joined a mainstem river system. The flow direction of this ancient mainstem river is unresolved. It is possible that it originated in the Challis and Absaroka volcanic fields and Wind River Mountains, flowed southeastward across southwestern Wyoming, and then veered eastward across the south flank of the Rock Springs uplift. Alternatively, the river could have originated along the western margin of the Sierra Madre Mountains in Wyoming and flowed westward across the south flank of the Rock Springs uplift toward the Bridger Basin. In either case, the lack of basement clasts is compatible with the interpretation that it was not regionally integrated.

Detrital sanidine and zircon data show that the Bishop Conglomerate and Browns Park Formation contain significant quantities of reworked Cenozoic tephra. Comparisons between the ages of young (younger than 40 Ma) detrital sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ and zircon U–Pb age distributions and previously published radiometric age and chemical data for western North America Cenozoic ignimbrites support the interpretation that the southern Basin and Range volcanic field was an important source of Bishop Conglomerate tephra. By comparison, the Southern Rocky Mountain, and probably the Snake River Plain, volcanic fields were important sources of Browns Park Formation tephra. The growing recognition that detrital zircon grains can originate as far-traveled volcanic ash fall has important implications for provenance interpretations based on detrital mineralogy.

Detrital zircon MDAs constrain the timing of collapse of the eastern Uinta Mountains, which was part of a more regional episode of extensional tectonism. Browns Park Formation sediment accumulation in the Browns Park graben marked the onset of collapse ca. 28–25 Ma. The abundance of young (12–8 Ma) tuffaceous Browns Park sediment in westernmost Browns Park suggests that this subregion had a later phase of collapse (post-12 Ma), which followed the initial structural foundering. This final phase of collapse extended the east-west-trending Browns Park graben farther west, and likely facilitated the establishment of the ancestral Green River, which eventually was integrated across the Uinta Mountains through Lodore Canyon after 8 Ma.

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