CRevolution 2—Origin and Evolution of the Colorado River System, Workshop Abstracts

May 24–26, 2010, U.S. Geological Survey, Flagstaff, Arizona

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CRevolution 2—Origin and Evolution of the Colorado River System, Workshop Abstracts

L. Sue Beard, Karl E. Karlstrom, Richard A. Young, and George H. Billingsley

Abstract

A 2010 Colorado River symposium, held in Flagstaff, Arizona, involved 70 participants who engaged in intense debate about the origin and evolution of the Colorado River system. This symposium, built upon two previous decadal scientific meetings, focused on forging scientific consensus, where possible, while articulating continued controversies regarding the Cenozoic evolution of the Colorado River System and the landscapes of the Colorado Plateau–Rocky Mountain region that it drains. New developments involved hypotheses that Neogene mantle flow is driving plateau tilting and differential uplift and new and controversial hypotheses for the pre-6 Ma presence and evolution of ancestral rivers that may be important in the history and birth of the present Colorado River. There is a consensus that plateau tilt and uplift models must be tested with multidisciplinary studies involving differential incision studies and additional geochronology and thermochronology to determine the relative importance of tectonic and geomorphic forces that shape the spectacular landscapes of the Colorado Plateau, Arizona and region. In addition to the scientific goals, the meeting participants emphasized the iconic status of Grand Canyon for geosciences and the importance of good communication between the research community, the geoscience education/interpretation community, the public, and the media. Building on a century-long tradition, this region still provides a globally important natural laboratory for studies of the interactions of erosion and tectonism in shaping the landscape of elevated plateaus.

Introduction

This report presents a summary report and abstracts from the CRevolution 2 Workshop—Origin and Evolution of the Colorado River System II, held at the U.S. Geological Survey (USGS) office in Flagstaff, Arizona, from May 24–26, 2010. The 70 registered participants (appendix A) were invited to submit abstracts, made available through a Google site for participants to share prior to the meeting. The agenda of the meeting (appendix B) preceded from the Gulf of California, up the lower Colorado River system, through Grand Canyon, across the central Colorado Plateau, to the Rocky Mountain headwaters and included studies in Arizona, California, Colorado, New Mexico, Wyoming, and Utah. Each speaker was allowed five minutes to present a few salient points, followed by five minutes of questions and discussion. This format of short, informal presentations provided fast-moving and lively sessions. In addition, many participants displayed posters at the meeting. The breadth of expertise and range of research was impressive and important in showing how the research community is excited and actively engaged in research on the history of the Colorado River system.

This report includes:

1) *Summary Report:* This section first presents the workshop in the context of previous research and meetings held in northern Arizona focused on the origin and evolution of the Colorado River, especially in context of the mystery of cutting Grand Canyon. Second, the report details both consensus on and continued controversy about certain key topics, as captured during vigorous discussions between the workshop participants. Finally, new developments and future research directions identified by the participants are outlined.

2) *Abstracts:* Many participants elected to revise their abstracts or submit new abstracts for publication in this open-file report. The abstracts range from one to as many as ten pages and are organized alphabetically.

3) *Appendixes:* Two appendixes are included: A, list of workshop attendees; B, workshop agenda.

Summary Report

CRevolution 2—Origin and Evolution of the Colorado River System

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Background

Studies of the origin and evolution of the Colorado River system are central to understanding the Cenozoic tectonic and geomorphic evolution of the western United States orogenic plateau and, more generally, to processes related to plateau formation and the development and integration of large river systems in complex tectonic regions. This region was uplifted from sea level in late Cretaceous to present elevations that exceed 4 km in the Rocky Mountains and 1.5 km over large regions of the Colorado Plateau. The Colorado River is the trunk of several tributary river systems that drain the western slope of the Rocky Mountains and the entire Colorado Plateau, and it is central to understanding the uplift and erosional history of the region. The Colorado River also has been recognized recently as having a major effect on not just the sedimentation, but the crustal evolution of the Pacific-North America plate boundary in the Gulf of California-Salton trough region (Dorsey, 2010; this meeting).

The timing of the initial development of the Colorado River, and its evolution into the drainage network seen today, has been the focus of more than a century of research, beginning with the early scientific trips of J.W. Powell down the Green and Colorado River systems. This field laboratory, because of its spectacular exposure, has been at the forefront of scientific breakthroughs in geomorphology, stratigraphy, paleontology, and tectonics (Dutton, 1882).

In early syntheses (Powell, 1879; Dutton, 1882), the premodern Colorado River system was treated as analogous to the present west-flowing river system that carries water from the Rocky Mountains to the Pacific. Longwell (1928, p. 143) noted the problem ("Muddy Creek problem") that the Colorado River did not exit at the western edge of the Colorado Plateau during the Pliocene, the lower boundary of which was placed at about 11 Ma until the early 1970s. Blackwelder (1934, p. 558–560) proposed that the regional river and canyon system did not exist until the Pleistocene, before which there were no integrated river systems. Hunt (1956) outlined the evolution of the entire region since Cretaceous time, and his Colorado River synthesis (Hunt, 1969) involved interacting geomorphic and structural controls on Colorado Plateau drainages through time.

Summarizing the continuing debate, Hunt (1969, p. 63) said: "The view that the Colorado River is an ancient river considers the river as a whole from the time of first uplift of the present Rocky

Mountains; the view that the river is young is based on particular segments." This was a glimpse of subsequent controversies surrounding attempts to reconstruct the regional picture by study of both regional uplift history and individual segments of the river system. Additional advances in our understanding of the complexities of the river system have been punctuated by three collaborative meetings in northern Arizona, in 1964, 2000, and 2010. This paper provides brief reflections on the first two meetings and a summary of the 2010 meeting. Our goal is to foster continued research on evolution of the western United States landscape at all scales.

1964 Meeting—Museum of Northern Arizona Colorado River Symposium

The first meeting had 21 participants. It was an outgrowth of discussions between Eddie McKee and Dick Young during visits to McKee's U.S. Geological Survey (USGS) office in Denver related to Young's Ph.D. dissertation (Young, 1966), funded in part by the Museum of Northern Arizona (MNA), Flagstaff. Young's fieldwork on the Hualapai Reservation and Ivo Lucchitta's work in the Lake Mead region (Lucchitta, 1966, 1972) evolved with close interaction. New data and participation of the two doctoral students was a major focus of the 1964 meeting. The symposium began with a three-day field trip to the Lake Mead country, then to Milkweed and Peach Springs canyons in western Grand Canyon. This was followed by the formal group discussions at the MNA for nearly a week. No formal talks, and very few slides, were allowed. Most of the data provided by the participants, other than Young and Lucchitta, had been published previously, and information from the symposium ultimately was integrated by McKee and others (1967). USGS geologist, Gene Shoemaker attended sporadically due to his urgent Apollo Project commitments, but he was a dynamic and influential force during the formal discussions and the field trips. Charlie Hunt, in spite of his influential works (Hunt, 1956) did not attend the 1964 symposium.

In retrospect, two things stand out about the 1964 symposium. First, some senior geologists (especially McKee, Shoemaker, and Koons) refused to readily accept the idea that the paleodrainage channels on the Hualapai Plateau that converge on Peach Springs Canyon flowed northeast and exited to the north across the course of the modern western Grand Canyon. They maintained their opinions despite the undeniable field data from gravel imbrication and clast lithologies. The conceptual problem was that these deep Tertiary canyons seemed to them to be heading into a "deep hole" from which there was no obvious outlet at appropriate elevations on the northern and eastern sides of Grand Canyon. The concept of northeast tilting of the Plateau to solve this issue of gradients (Young, 1982) was not strongly argued until after the report by McKee and others (1967). There also was hesitancy to accept the idea that the oldest basal arkosic gravels along the edge of the plateau (Rim gravels) could be older than Miocene (now known to be Paleogene or late Cretaceous; Young and Hartman, this volume). The subsequent McKee and McKee (1972) article on Pliocene uplift of the Colorado Plateau that was written to explain the Rim gravels, attests to the difficulty of changing minds about the history of the Plateau margin, despite the knowledge by then that the Peach Springs Tuff that caps the gravels in the Hualapai Plateau sections was 18.5 Ma (Young and Brennan, 1974) and that the Pliocene-Miocene boundary at that time had recently been moved from 11 Ma to 5 Ma.

Second, there were few individuals at the meeting who knew details of the little-studied Cenozoic history of the Little Colorado River Valley and its environs. Therefore, despite the perceived young age of the Bidahochi Formation deposits (~4–6 Ma age at the time; McKee and others, 1967), it was decided to "send" the Colorado River off to the south (by default), presumably accompanying ponding of the drainage in Lake Bidahochi. There seemed to be no other place for the ancestral river to go. Bidahochi ages have since been revised to 16–6 Ma, and an alternate southern escape route through the ancestral Salt River Canyon has been resurrected by Potochnik (this volume). Thus, much of the uncertainty concerning the timing of events as perceived in 1964 needs to be put in the context of the relative lack of accurate ages of key Cenozoic units.

Nevertheless, the main accomplishment of the 1964 symposium was to outline the state of knowledge for different parts of the Colorado Plateau more clearly and to combine the ideas of major researchers (other than those of Charlie Hunt). McKee and others (1967) summarized some of the main questions about Colorado River evolution: (1) time of initiation, (2) processes of integration, and (3) early paleodrainage courses. They emphasized river segments that may have had different earlier histories and been integrated into the Colorado River system that we see today after Muddy Creek time. Then, as now, there was little consensus about pre-6-Ma river geometries, but the stage was set for continued debate.

2000 Grand Canyon Meeting—Colorado River, Origin and Evolution

This meeting had 73 formal registrants and was held at Grand Canyon National Park in June 2000 (Young and Spammer, 2001). By the time of the 2000 meeting, the maturation and application of plate-tectonics concepts, much more field data, and many more K-Ar ages had dramatically improved the understanding of chronology and sequence and timing of events. Yet the central problems of the location of a postulated Miocene river and how Colorado River integration occurred remained unresolved.

The meeting and resulting collection of papers was an outgrowth of informal conversations among Colorado Plateau geologists during the 1990s. The purpose of the symposium was to review knowledge of the geologic issues, controversies, and progress regarding the geologic evolution of the Colorado Plateau and the Colorado River. The meeting (June 5–11, 2000) was coordinated by R.A. Young, with significant input from George Billingsley. Grand Canyon National Park superintendent Robert L. Arnberger underwrote the major expenses for the meeting, with other financial and informal donations of materials and personnel time provided by the USGS Flagstaff office (George Billingsley, Sue Beard, Sue Priest), Grand Canyon Association (Greer Price), SUNY Geneseo Department of Geological Sciences (R.A. Young), Northern Arizona Uuniversity Departments of Geology and Geography (Michael Ort, Lee Dexter), the Arizona Geological Survey (Jon Spencer), and the Nevada Bureau of Mines (James Faulds). The meeting was coordinated by Greer Price, with assistance from Tom Pittinger of the National Park Service for accommodations, meals, and meeting facilities. Field trips were led by Michael Ort to view the Bidahochi Formation stratigraphy in eastern Arizona, and Prescott College professor, Andre Potochnik, to view Mogollon Rim geology, with a postmeeting field trip to view the Tertiary geology of the Hualapai Indian Reservation and Lake Mead led by Richard Young, James Faulds, Sue Beard, Keith Howard, and Ivo Lucchitta.

Many papers reported on new dating techniques applied to elevated terraces of the river system to infer incision rates. The emerging picture was a river with both spatially and temporally varying incision rates along its course, with differential incision rates related to both geomorphic and structural controls. The presence of a well-integrated, ancestral upper Colorado River drainage system in Colorado and southern Utah, as postulated by Hunt (1969), was not strongly supported. Debate continued about where ancestral upper Colorado River water and sediment loads were stored before integration at the mouth of Grand Canyon ~4–5 Ma. The Bidahochi Formation, as evidence for such a Miocene lake, seemed acceptable from a chronologic perspective, but not necessarily from a sedimentological viewpoint. Timing of Colorado Plateau uplift(s) remained controversial, with advocates for both late Tertiary uplift and Laramide (~70–50 Ma) uplift. Late Pleistocene incision rates were reported to be rapid enough to carve Grand Canyon within the last 10 Ma, but this raised the significant issue of why rapid incision of the entire basin did not begin and progress more rapidly immediately following Miocene Basin and Range extension, when appreciable relief developed between the Colorado Plateau and the extended terrane to the west. No consensus was reached about mechanisms by which different river segments may have been integrated. Both lake spillover and headward erosion models were advanced again, and other controversies were aired. (1) When did canyon cutting first begin? (2) Which way were rivers flowing in early Tertiary time? (3) How much Mesozoic and Cenozoic sediment overlay the Permian Kaibab Formation in different areas, and how fast did erosion denude the landscape? (4) How did 5–6 Ma lake systems, represented by the Bidahochi, Hualapai, and Bouse deposits, relate to a through-going Colorado River?

The 2000 meeting marked renewed interest in, and progress on, all aspects of the Cenozoic evolution of the Colorado River system. A plethora of models were discussed, and the meeting seemed to mark an attempt to compile all objective criteria (separate from a model-driven approach) for the timing of the development of various surfaces and paleosurfaces, the rate and timing of cooling of rocks as they were unroofed towards the modern surface, and rates of river incision through time. The meeting and resulting volume catalyzed renewed research and the integration of diverse scientific approaches, all aimed at resolving the landscape evolution of the Colorado Plateau/Grand Canyon region. It also spawned popular treatments of longstanding controversies about evolution of Grand Canyon (Powell, 2005; Ranney, 2005).

Continued challenges and questions were identified at the meeting. (1) What were the causes and precise timing of plateau uplift(s)? (2) How much Mesozoic and Cenozoic sediment was deposited in the Grand Canyon region, and when did it get stripped off? (3) Could a western Grand Canyon precursor stream have been present without leaving a preserved sedimentary record near the present mouth of Grand Canyon? (4) Did integration across the Kaibab uplift take place by headward erosion or basin spillover? (5) What role did local or global climate change play in enhancing or delaying the incision and integration of the Colorado River system?

2010 Flagstaff Meeting—CRevolution 2—Origin and Evolution of the Colorado River System II

The 2010 meeting follows in the footsteps of prior meetings in several important respects. It represents an assembly of many of the key scientists and their students researching the evolution of the Colorado River system. This meeting followed a comprehensive regional approach (for example, Hunt, 1956) involving detailed studies from the Gulf of California to the high Colorado

Rocky Mountains, including application of new geochronologic and analytical techniques to quantify rates and model processes of landscape evolution. New aspects of this meeting were: (1) the examination of links between mantle processes and their potential surface effects; (2) discussion and quantification of the isostatic response to denudation that affects landscapes; (3) increased emphasis and emerging syntheses of low-T thermochronology (apatite-fission track and apatite-helium studies); (4) discussion of groundwater sapping as an important river-integration mechanism; and (5) greater emphasis on process-oriented studies, all aimed at understanding driving mechanisms, timing, and magnitudes of differential river incision and landscape denudation and their tectonic connections.

Table 1 lists the abstracts presented at the meeting; the abstracts are referenced in the following discussion. The agenda of the meeting proceeded from the Gulf of Mexico, up the lower Colorado River system to Grand Canyon, and across the central Colorado Plateau to the Rocky Mountain headwaters. Like the 2000 workshop, the 2010 workshop reinvigorated researchers regarding the Colorado River region in the context of regional and global questions about tectonic and geomorphic processes that shape landscapes.

Invitees submitted extended abstracts to an Internet site so that all participants could access and read these informal contributions before the meeting. The format of the workshop was designed to encourage discussion and data compilation in a format different from the formal talks presented at most professional meetings. Oral remarks were limited to 5 minutes and were followed by extensive plenary discussion among the participants. Most abstracts presented at the meeting were revised and reviewed for this open-file report. Manuscripts; electronic databases, such as geochronology; incision data; and useful maps and images of the Colorado River system developed for this meeting will be submitted as separate contributions to a future Geosphere volume that has been proposed.

Toward Consensus

The meeting moved toward consensus on several topics.

(1) Multiple episodes of erosion and uplift. Regional geologic and thermochronologic data (Kelley and others; Lee and others) indicate that punctuated episodes of erosion and inferred uplift took place in Laramide time (Wernicke; Lee and others; Young and Hartman), middle Tertiary time (Cather; Lee and others), and during the last 10 million years (Karlstrom and others; Hoffman and others). Debate continues regarding the duration and nature of different tectonic and(or) climatic forces, and which episode was dominant in a given region or reach of the river system.

(2) Drainage reversal(s). The concept of drainage reversal on the Colorado Plateau seems well established. Ancestral rivers flowed north (Davis and others; Hill and others), or northeast and east (Wernicke; Potochnik) during Late Cretaceous (Wernicke) to Paleocene-Eocene time (Davis and others; Young and Hartman; Young and Crow; Beard and Faulds), whereas the post-6 Ma Colorado River flows southwest. Debate continues about the timing and drainage geometry of most of the pre-6 Ma paleorivers and mechanisms driving drainage reversals.

(3) Mid-Tertiary erosion. Middle Tertiary time, after the Oligocene Chuska erg of the Four Corners region (Cather), represented a time of regional deep erosion on parts of the Colorado Plateau that is documented by ~25 Ma cooling based on thermochronology data in

Grand Canyon and the Colorado Rocky Mountains (Kelley and others; Lee and others). Tectonic influences on this denudation are debated.

(4) Age of the upper Colorado River system. Oligocene drainages systems are documented by gravels beneath 25 Ma basalts (Aslan and others) and west-draining Oligocene paleocanyons in the Gunnison Colorado region (Sandoval and others). By 10 Ma, evidence of a paleo-Colorado River in the Colorado Rockies is seen in gravels beneath the 10.4 Ma Grand Mesa basalt (Aslan and others, Cole) and several other ~10 Ma basalts (Lazear and others). Onset of rapid incision and denudation in the upper Colorado River paleodrainages took place between 10 and 6 Ma, as documented by thermochronology from a deep well near Rifle, Colorado (Karlstrom and others).

(5) Age of the Lower Colorado River system. The 5–6-Ma age of integration of the Colorado River system, from the Rocky Mountains across the Kaibab Plateau to the Gulf of California, continues to be supported by data from the 5.3-Ma age of the first Colorado Plateau-derived sediments arriving in the Gulf (Dorsey; Kimbrough and others), lack of Colorado River sediments in the Grand Wash trough (Muddy Creek constraint; Lucchitta), and geometry of late Miocene alluvial fans now dissected by the Colorado River and Grand Canyon (Lucchitta and others). Sedimentary budgets suggests that the volume of sediment in sedimentary basins of southern California is roughly compatible with estimates for erosion of material off the Colorado Plateau in the last 6 Ma (Dorsey), consistent with post-6-Ma integration. Debate continues about the downstream course and role in river evolution of pre-6-Ma paleocanyons that may have become reactivated and linked to evolve into the modern Grand Canyon (Young, 2008).

(6) Lake spillover along the lower Colorado River. Lake-spillover models for the lower Colorado River (House and others; Howard) are increasingly well-documented by mapping of Pliocene deposits in the Lake Mohave area, and support continues for a lacustrine origin for the Bouse Formation in the Mohave Valley through Parker Valley region (House and others; Malmon and others). Contrary to some older models suggesting a marine origin, Sr, O, and C isotopic data support a lacustrine origin for the upper Bouse in the Lake Mojave area and Hualapai Limestone near the mouth of the Grand Canyon (Spencer and others; Crossey and others; Lopez Pierce and others).

(7) Bullhead aggradation. Major aggradation in the lower Colorado River at about 5.5–3.3 Ma is well-documented by the Bullhead alluvium and related gravels (House and others; Howard and others), although explanations for this event, and for the steep river profile (as much as twice that of the modern river) for the aggradational sequence (Howard), are debated.

(8) Integration Processes. In general, all river-integration processes of lake spillover, headward erosion, and groundwater sapping (Crossey and others; Pederson; Hill and others) were considered viable (Douglas) and may have operated in combination. Although the 5–6 Ma timing of integration generally is agreed upon, the importance of these mechanisms for integration of the Colorado River system remains controversial.

(9) Differential incision. Evidence is accumulating for different incision rates through time and space in the river system, and patterns of differential incision are becoming better resolved by combined geochronologic and geomorphic data (Crow and others; Marchetti and others). Where differential incision can be shown to be related to fault dampening of incision, as in western Grand Canyon (Crow and others), it is indicative of a dynamically changing river system adjusting to tectonic forces. The relative importance of geomorphic, climatic, and tectonic controls on drainage evolution are important issues being debated.

(10) Rapid onset of denudation 6–10 Ma. There also is improved evidence from apatite-helium retention ages, as well as geologic studies for regional acceleration of exhumation and incision in Neogene time. This occurred after about 5–7 Ma in upper Grand Canyon (Lee and others), Little Colorado River (Embid and others), Monument Uplift, Canyonlands, and Roan Cliffs (Hoffman and others). Incision accelerated starting 6–10 Ma in the Grand Mesa, Colorado area (Aslan and others; Cole; Karlstrom and others). Debates continue about the extent to which this was driven by tectonic uplift (Karlstrom and others) or combinations of drainage integration (Pederson), enhanced Pleistocene runoff, the Southwest monsoon climate, and the opening of the Gulf of California in the Miocene (Hoffman and others)

(11) Isostatic response to denudation. Isostatic consequences of erosion involve rebound of the crust to balance the load removed (Pederson; Lazear and others). Faulting and differential erosion during possible tilting also have isostatic responses. Quantification of this component of landscape evolution is important and is being studied by several groups.

(12) Paleogeography reconstructions. Because of growing evidence for post-Miocene tectonic and isostatic adjustments to surface elevation, we cannot rely solely on modern elevations to reconstruct past elevations and geometries of paleolake shorelines, spill over points, and paleoriver gradients. The details of these structural and isostatic adjustments are a growing area of research.

(13) Mantle-driven uplift. New geophysical images that show large contrasts in mantle velocity (and inferentially temperature and rheology), over <100 km spatial scales, provide strong evidence for Neogene mantle flow and tectonism in the western United States In addition, chemical and isotopic data indicate both asthenospheric and lithospheric sources for Neogene basalts (Karlstrom and others). Geodynamic models suggest that observed mantle-velocity variation should drive surface uplift and subsidence (Karlstrom and others; Robert and others), but debate continues about timing and nature of mantle flow. Several geodynamic models suggest that the magnitude of predicted effects on surface topography is on the order of 400–800 m of uplift.

Continued Controversies

Many of the controversies discussed during the 1964 and 2000 symposia persist.

(1) Opening of the Gulf of California. While most published data support a latest Miocene age (~6.5 Ma) for initial marine incursion (Dorsey), paleontological data support marine conditions starting in middle Miocene time (McDougall). Middle Miocene opening of the Gulf would suggest that it did not play a major role in the integration of the Colorado River around 5.5–6 Ma. Top-down (for example, lake spill-over) integration models also do not rely on opening the Gulf of California to lower base level and directly trigger integration, although the Gulf opening may have intensified summer monsoons and, therefore, erosion rates (Wernicke).

(2) The Bouse Formation. Marine versus nonmarine origin for the lowermost Bouse Formation along the Colorado River in the southern Yuma Basin was debated again in 2010 (McDougall). Some suggested that the Bouse Formation records a change from a marine environment in the Yuma Basin to nonmarine conditions in the northern Mojave paleolake. Sr, O, and C isotopic data from lower Bouse carbonates are consistent with mixing trends between river, marine, and deep bedrock sources of water for parts of the Yuma Basin; Sr isotopes alone do not provide conclusive evidence for nonmarine origin (Crossey and others). In contrast, others point towards the similar character of limestones in the basal Bouse Formation in all areas, and

nonmarine isotopic signatures (Spencer and others), to support a nonmarine origin in all of the subbasins.

(3) Comparisons between the Green and Colorado Rivers. Two great rivers converge in Canyonlands to form the Colorado River system. The Colorado River is steeper and has higher incision rates during the last few million years than the Green River (Aslan). Controversies arise concerning when the Green became established as a south-flowing river (Ferguson, Pederson) and became integrated with drainage from the Colorado Rocky Mountains, and whether the different gradients reflect differential uplift of the Colorado Rocky Mountains relative to the central Colorado Plateau (Karlstrom and others). Young denudation also is evident in Wyoming based on thermochronologic and stratigraphic data, and it is suggested that the Green River switched its flow direction from north to south in the latest Miocene (Ferguson).

(4) Where did Miocene paleorivers flow? The longstanding question of where Miocene upper Colorado paleorivers may have exited, or terminated within, the Colorado Plateau is unresolved. One model proposes internal drainage in the western Rocky Mountains until 6 Ma, separated by the Kaibab uplift from west-flowing Miocene paleorivers in western Grand Canyon (Wernicke; Hill and others; Lopez Pearce and others). In contrast, evidence from gravels in Wyoming suggests a possible north-flowing system in Miocene time (Ferguson). Alternatively, the idea of a southerly exit, along the Salt River system, was revived (Potochnik).

(5) Pre-6 Ma paleocanyons and paleorivers. Numerous workers have proposed models by which pre-6 Ma paleocanyons on the southern Colorado Plateau may have become re-occupied and linked to evolve into the modern Grand Canyon. A possible west-flowing Miocene river preserved along Crooked Ridge (Lucchitta and others) may have fed into a system occupying the present location of eastern Grand Canyon (Lee and others; Pederson) or the entire Grand Canyon (Wernicke). However, the latter model, especially, is in conflict with the Muddy Creek constraint (Lucchitta). Geologic evidence argues against the presence of at least some of the proposed paleocanyons. For example, a precursor western Grand Canyon drainage (Hill and others; Wernicke; Young; and Cole) seems to be negated by the absence of zircons indicative of Paleozoic sedimentary sources in detrital zircon populations from 13–6 Ma rocks of the Muddy Creek Formation near Pearce Ferry, suggesting these deposits could not have had detrital input from the Paleozoic strata east of the western Grand Canyon (Lopez Pierce and others). It was noted that the Virgin depression and northern Grand Wash are deep extensional basins based on geophysical data that may have received detritus from Miocene rivers flowing to the northwest across the Colorado Plateau (Umhoefer and others).

(6) An old Grand Canyon. The possibility of a Late Cretaceous (70 Ma) paleocanyon, coincident with both the eastern and westernmost segments of the modern Grand Canyon and cut to within 400 m of its present depth, was supported by a new interpretation of published thermochronology data (Wernicke). This was hotly debated by both thermochronologists (Kelley and others; Lee and others) and geologists (Karlstrom and others) and provides a hypothesis that challenges other models.

(7) Integration mechanisms. Possible mechanisms of integration of the upper and lower Colorado River basins across the Kaibab uplift include lake spillover, piracy, headward erosion, and groundwater sapping through karst connections. Different types of proposed groundwater and karst connections included: (1) ~6 Ma karst-piping of river waters from the upper basin under the Kaibab uplift (Hill and others); (2) upper-basin seepage through paleocanyons and karst to an integration point in the central-western Grand Canyon region (Pederson); and (3) groundwater sapping from locally sourced groundwater (not upper-basin river water), where

hydrologic head facilitated incision and integration, while geochemical signals of local groundwater were preserved in spring-discharge deposits (Crossey and others). Simple headward erosion as a dominant integration mechanism was supported by some (Hill and others), as was piracy and integration of existing drainage systems by a top-down overflow process (Douglass). The striking similarity of detrital zircon populations in the modern river delta to 5.3 and 4.4 Ma Colorado River deposits (Kimbrough and others) suggests a top-down integration because headward erosion would predict progressive changes in detrital populations through time in the river's lower reaches and delta, which are not observed. Furthermore, thermochronology data that indicate rapid onset of denudation at about the same time in several places across the region (Hoffman and others) are hard to reconcile with headward-erosion models.

(8) Lake Bidahochi. The size and significance of a hypothesized paleo "Hopi Lake", or "Lake Bidahochi", and the depositional setting for the Bidahochi Formation were debated, and several models were presented. (1) This Miocene basin was a terminal, internally drained depression for southward-flowing waters from the southern Rocky Mountains. (2) This lake system may have been a headwater lake for a regional northward-flowing river that carried Rocky Mountain drainage into Wyoming (Ferguson), or drained into other hypothetical lakes near Lees Ferry (Hill and others). Models for integration of the Colorado River driven by spillover from Lake Bidahochi were not strongly supported by facies analysis of the Bidahochi Formation, which suggests low sediment-accumulation rates in a small lake (Pederson), where fluvial beds aggraded across a more limited lacustrine facies (Dickinson).

(9) Drainage reversal. The concept of drainage reversal from Paleocene to Eocene north- to east-flowing systems (including the paleo Salt River), to the post-6-Ma southwest-flowing Colorado River system is better constrained (Young and Hartman), but much discussion centered on the timing and mechanisms for this reversal. Tilting due to mantle-driven epeirogeny (Robert and others; Karlstrom and others) was suggested as a mechanism of drainage reversal, as well as a possible driving force for river integration and propagation of knickpoints (Darling and others).

(10) Nature of knickpoints. The cause and significance of knickpoints and convexities in longitudinal profiles of the Colorado River and its tributaries were debated as either (1) being fixed at less-erodable rock layers or reaches, as documented by studies of bedrock-strength properties along the river profile (Tressler and others); or (2) incision transients propagating upstream in response to downstream tectonic and(or) geomorphic (for example, piracy) events (Darling and others). Diffuse knickpoint migration to bypass a bedrock obstruction may help explain high incision rates above Lees Ferry (Hanks and others; Marchetti and others; Pederson). Mantle tomographic images suggest the Lees Ferry knickpoint is caused by dynamic forces owing to mantle flow associated with a pronounced mantle velocity gradient (Karlstrom and others) that may help explain differential incision rates above and below Lees Ferry (Darling and others)

(11) Isostatic response to denudation. The relative roles of tectonic uplift and isostatic responses to denudation to drive rock uplift were discussed to explain differences in incision rates. According to one model (Pederson and others), calculated magnitudes of isostatic response to erosion correlate with the pattern of faster Pleistocene incision rates in the central Colorado Plateau, suggesting that isostatic feedback accounts for much of those amplified rates. A second model (Lazear and others) suggested that the difference between river incision rates and calculated isostatic response to denudation during the last 10 m.y. in the Grand Canyon and

Colorado Rocky Mountains requires Neogene tectonic-uplift components at both ends of the river system.

(12) Timing and mechanisms of uplift. Timing and the process of uplift of the surface of the Colorado Plateau and southern Rocky Mountains, from sea level in late Cretaceous time to modern high elevations, and the interactions of uplift, drainage development, and erosion, remain essential questions. Paleoelevation data from clumped isotopic analysis suggests that most uplift in the southwestern Colorado Plateau was accomplished in Laramide time, which pairs with "old canyon" models (Hill and others; Wernicke). Thermochronology indicates Miocene cooling in eastern Grand Canyon (Lee and others) about the same time as broad denudation across the southern plateau (Cather). This timing is consistent with various proposed mechanisms, including lithosphere delamination, conductive mantle heating, Farallon slab removal, and whole mantle flow (Robert and others). Evidence for Neogene and ongoing mantle flow and resulting uplift can be paired with young canyon models (Karlstrom and others). If onset of rapid denudation on the Colorado Plateau predated Colorado River integration ~6 Ma, recent surface uplift and tilting seem to be required (Karlstrom and others). Alternatively, if onset of rapid denudation postdated integration, tectonic uplift would not be required to create a young Grand Canyon (Hoffman and others).

New Developments and Future Research Directions

Directions for further research involve application of new methodologies and better integration of diverse datasets:

(1) Continue detrital zircon studies of paleo-Colorado River deposits along the mainstem and of all tributary deposits to help resolve the processes of integration and evolution of the Colorado River system.

(2) Develop and test paleobarometers to estimate absolute elevation changes through time and thus, investigate links between topographic changes and surface-uplift events.

(3) Evaluate the timing and locations of drainage reversals by studying the ages of terrace gravels and lake deposits.

(4) Study the highest terraces of the Green and Colorado Rivers, and the Browns Park Formation, southern Wyoming, in order to better document how and when the Colorado and Green Rivers became integrated, and whether steeper gradients of the upper Colorado River are due to rock uplift of the southern Rocky Mountains.

(5) Reconcile apatite fission-track ages and U-Th-He ages with one another and with other geologic constraints. Both techniques should be applied routinely to the same samples.

(6) Produce interlaboratory protocols and reduce uncertainty in how to interpret variable apatite ages from the same sample.

(7) Apply thermal models to reduce uncertainty in estimating timing of onset of rapid denudation in diffuse age-elevation plots. Reducing the uncertainty could help, for example, to resolve whether onset of rapid denudation in the Colorado Plateau areas was pre-6 Ma, in which case river-integration is not the causative explanation, or syn- to post-6 Ma, in which case a river integration model might predict an upstream-younging of onset of rapid denudation (so far not observed).

(8) Integrate sediment-budget studies with thermochronology in order to establish links between upland denudation events and downstream sediment volumes and major aggradation events.

(9) Evaluate evolution of river profiles through time by using incision-rate studies at all temporal and spatial scales. Precise dates, definitive strath heights, and, ideally, depth to bedrock are needed to calculate reliable bedrock-incision points. Strath-to-strath comparisons for a given reach are especially valuable in calculating bedrock-incision rates independent of depth to bedrock in the river channel. Strath-to-strath age data are also needed to test models of changing incision rates through time versus steady-rate incision models.

(10) Complete structural studies and models (for example, Resor) that provide better understanding of fault-slip history, monoclinal-fold formation, and possible eperiogenic doming or tilting. Furthermore, structural studies should be better integrated with incision and cooling/denudation (thermochronology) interpretations. Locations need to be sought (for example, Lees Ferry and Grand Mesa regions) where long-term incision-rate data can be merged with age-elevation lowtemperature-thermochronology data in order to create integrated denudation-incision-rate databases (in m/Ma).

(11) Test geodynamic models of mantle flow against improved differential-incision and differential-denudation models.

(12) Test geomorphic models of knickpoint migration against improved differential-incision and differential-denudation data.

(13) Produce improved community-based databases on geochronologic, incision-rate, and thermochronologic constraints for research on the evolution of the Colorado River system that could be continually updated from new research.

(14) Build on graphic visualizations that evolved between the 1964, 2000, and 2010 meetings by creating GIS-based paleogeographic maps that are tied to a detailed timeline and spatially referenced to up-to-date comprehensive databases. This could lead to time-lapse sequence movies of the evolving landscape tied to evolving lithospheric structure.

Outreach

In addition to scientific goals, meeting participants emphasized the iconic status of Grand Canyon for geosciences and the importance of good communication between the research community, the geoscience education/interpretation community, the public, and the media. About 5 million visitors come to Grand Canyon each year, and most want to know how old it is and about the processes that shaped it. The research community has an important obligation to convey new research advances and educational resources involving the spectacular landscapes and field laboratories of the Colorado Plateau-Rocky Mountain region. Informal research meetings of the type done here provide exceptional research and public-relations value for the geosciences.

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Table 1. Abstracts presented at the 2010 meeting. (Note: Not all abstracts presented were submitted to this volume.)
Amoroso, L.,Felger, T., and Wan, E., The Willow Beach beds—A pre-Colorado River axial-basin deposit.
Aslan, A., and the CREST Working Group, Origin of the ancestral Colorado and Gunnison Rivers and post-10 Ma river incision rates in western Colorado.
Beard, L.S., and Faulds, J.E., Kingman uplift, paleovalleys, and extensional foundering in northwest Arizona.
Blakey, R.C., Ranney, W., and Loseke, T., Oligocene-Early Miocene incision, strike-valley development, and aggradation, Mogollon rim, Verde Valley region, Arizona—A potential analogue for pre-Grand Canyon development.
Cather, S.M., Late Oligocene—early Miocene deep erosion on the southern Colorado Plateau and the southern Great Plains.
Cole, R.D., Significance of the Grand Mesa basalt field in western Colorado for defining the early history of the upper Colorado River.
Crossey, L.J., Karlstrom, K.E., Lopez Pearce, J., and Dorsey, R., Geochemistry of springs, travertines and lacustrine carbonates of the Grand Canyon region over the past 12 million years—The importance of groundwater on the evolution of the Colorado River system.
Crow, R., Karlstrom, K.E., and McIntosh, W., Incision history of Grand Canyon from dated Colorado River gravels.
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Davis, S.J., Dickinson, W.R., Gehrels, G.E., Spencer, J.E., Lawton, T.F., and Carroll, A.R., The Paleogene California River—Evidence of Mojave-Uinta paleodrainage from U-Pb ages of detrital zircons.
Dickinson, W.R., Bidahochi paleogeography and incision of the Grand Canyon.
Dorsey, R.J., A sediment budget for the Colorado River.
Douglass, J., One Grand Canyon but four mechanisms: Was it antecedence, superimposition, overflow, or piracy?
Embid, E.H., Crossey, L.J., and Karlstrom, K.E., Incision history of the Little Colorado River based on K-Ar dating of basalts and U-series dating of travertine in the Springerville area.
Felger, T.J., Fleck, R.J., and Beard, S.J., Miocene-Pliocene basalt flows on the east and west flanks of Wilson Ridge, Arizona, preserve multiple stages in the depositional history of adjacent Detrital Wash and Black Canyon Basins, and may help constrain timing of incision by the Colorado River.
Ferguson, C.A., Powder Rim gravel, deposit of a late Miocene, north-flowing river through the Wyoming-Colorado-Utah borderland.
Hanks, T., Blair, L., Cook, K., Davis, M., Davis, S., Finkel, B., Garvin, C., Heimsath, A., Lucchitta, I., Webb, B., Whipple, K., and Young, D., Incision rates of the Colorado River in Glen Canyon.
Hill, C., Ranney, W., and Buecher, B., A working model for the evolution of the Grand Canyon/Colorado Plateau Region—Laramide to present.
Hoffman, M., Stockli, D., Kelley, S., Pederson, J., and Lee, J., Mio-Pliocene erosional exhumation of the central Colorado Plateau, eastern Utah—New insights from apatite (U-Th)/He thermochronometry.
House, P.K., Pearthree, P.A., Brock, A.L., Bell, J.W., Ramelli, A.R., Faulds, J.E., and Howard, K.A., Robust geologic evidence for latest Miocene-earliest Pliocene river integration via lake spillover along the Lower Colorado River—Review and new data.
Howard, K., Pliocene aggradational sequence of the lower Colorado River in longitudinal profile.
Howard, K.A., and Malmon, D.V., Boulders deposited by Pliocene and Pleistocene floods on the lower Colorado River.
Howard, K., Malmon, D., McGeehin, J., and Martin, P., Holocene aggradation of the lower Colorado River in Mohave Valley, California and Arizona.
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Lucchitta, I., The Muddy Creek Formation at the mouth of the Grand Canyon: Constraint or chimera?
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Pederson, J., Drainage integration through Grand Canyon and the Uintas—Hunt and Hansen's groundwater-driven piracy via paleocanyons.
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The Willow Beach Beds—A Pre-Colorado River Axial-Basin Deposit

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The Colorado River below Lake Mead, between Arizona and Nevada, flows through a rugged landscape produced by late Miocene volcanism, extensional tectonism, and subsequent erosion. The interior-drained axial basin between Hoover Dam and Cottonwood Valley, informally called the Black Mountain Basin (Williams, 2003), was created within the Colorado River extensional corridor and filled with axial channel and fan sediments, evaporites, and carbonates from ~13–6 Ma (Faulds and others, 2001b). The basin deposits interfinger with fanglomerates and megabreccia deposits derived from the east. The Colorado River entered the basin 6–5 Ma; subsequent erosion has removed much of the central portion of the axial-basin deposits except where capped by resistant Malpais Flattop basalt or megabreccia. The mostly fine-grained deposits of uncertain age preserved along eastern side of the Colorado River at a prominent bend in the river are informally called the Willow Beach beds (fig. 1). The Willow Beach beds (Twb, fig. 1) are unconformable on Proterozoic crystalline rocks or, locally, on boulder fanglomerate (T3a) derived from Proterozoic and mid-Miocene volcanic rocks. The basal contact is exposed along parts of Jumbo Wash and the lower portions of some of the major washes that terminate at the river. The basal Willow Beach beds consist of alluvium that becomes finer from south to north. At the southern end of the Twb exposures south of the Willow Beach Road, the sediments consist of weakly to moderately lithified subangular cobble/gravel conglomerates, mostly matrix-supported, interbedded with thin to moderate-bedded medium to coarse-grained sandstones and thin-bedded fine to medium-grained sandstones. Imbrication measurements of the conglomerate clasts indicate the sediment source to the southeast and southwest.

Farther north along exposures in Disposal Wash (fig. 2), the basal sediments contain fewer gravels in weakly lithified medium- to coarse-grained sandstone. Where the Willow Beach beds are best exposed along Disposal Wash, the section shows basal conglomeratic sandstone grading upward to light reddish-brown bedded sandstones and light-brown clayey sandstones. A 9-m thick massive gypsum and gypsiferous sandstone unit about 70 m above the base indicates that the axial-basin contained either a playa or a brackish lake. Above the massive gypsum are well-bedded medium- to coarse-grained sandstones and fine to medium clayey sandstones. The upper 12 m of the exposure is reddish-yellow gravel to cobble sandstone and gravel-poor sandstone. Clasts are gneiss, felsic and mafic volcanics, and red granites. The contact with the overlying T2a fanglomerate appears gradational in exposures along Disposal Wash. The Willow Beach beds contain several thin-bedded tuffs and volcaniclastic beds along with rare pumice clasts. Many of the tuffs are discontinuous over 10–20 m, and range from 3 to 20 cm thick. The presence of pumice clasts up to 8 cm suggests a local source or a nearby source, such as the 16–7 Ma southwestern Nevada volcanic field (Perkins and others, 1998). The uppermost tuff bed of the Willow Beach beds shown in figure 2 was submitted for major and minor oxide analysis. Geochemical correlation of the uppermost tuff deposit by a coauthor yielded the best correlation with a locally derived but undated tephra from the vicinity of Oreana, Nevada. The Willow Beach bed tuff is chemically similar, but with a lower correlation coefficient, to tephras collected from Fish Lake Valley, California and Redlich Summit in west-central Nevada. These correlations suggest a preliminary age of 5.9–5.3 Ma for the tephra. However, Williams' (2003) mapping implies the Willow Beach beds are older.

The Black Mountain Basin is an asymmetric basin, probably a half-graben (Williams, 2003). The eastern margin of the basin is bounded by two down-to-the-west normal faults; the northern and southern parts of the basin are bounded by the Fortification Ridge fault (Mills, 1994) and the Bighorn fault (Faulds, 1993). There is likely a continuous west-dipping fault between these two normal faults, based on the dip of the Willow Beach beds, but its presence has not been confirmed (Faulds and others, 2001a). In the northern part of the basin, Williams (2003) observed 5–15º east-dipping sandstones and fanglomerates and fanning dips, suggesting, extension at the same time as material eroded from the Black Mountains was deposited in the basin. The Willow Beach beds dips 3–5º east throughout the section, implying deposition was contemporaneous with the latest extension.

Three options are proposed for correlation of the Willow Beach beds with similar deposits in nearby basins. The Willow Beach beds may correlate with Muddy Creek beds near Frenchman Mountain (Anderson, 1978; Castor and Faulds, 2001) or the Lost Cabin beds in the Cottonwood Valley (House and others, 2005). The basin near Frenchman Mountain contains gypsum, marl, limestone and reddish sandstone that Castor and Faulds (2001) correlated with the Muddy Creek Formation. The redbeds contained a vitric ash that was geochemically correlated with the 5.59 Ma tuff of Wolverine Creek erupted from the Heise volcanic field in the eastern Snake River Plain. The Lost Cabin beds are pre-Colorado River flat-lying interbedded sandstone and mudstone deposits that contain the 5.5 Ma Connant Creek ash (also from the Heise volcanic field), but contain no evaporites or carbonates. Alternatively, the Willow Beach beds may be related to nearby 13–11 Ma Tml deposits described by Williams (2003), shown in figure 2.

Williams (2003) described and measured sections of Tml, a part of the Black Canyon Assemblage, and used Anderson's (1978) Tmu and Tml nomenclature (fig. 2). Williams (2003) identified these deposits as being shed from the Wilson Ridge-Black Mountains area during extension and uplift. One of Williams' (2003) measured sections was located north of the basalt Tfb (north end of the map, fig. 1). This basalt, now informally called the Hwy 93 basalt, has been recently dated at 5.46±0.03 Ma (Felger and others, 2010). The second of Williams' measured sections was of the lower Black Mountain conglomerate in the Malpais Flattops area south of the Willow Beach bed outcrops (southern end of the map, fig. 1). Williams (2003) dated an ash layer ($^{40}Ar/^{39}Ar$) in Tml of the northern Black Canyon Assemblage at 11.72±0.06 Ma and geochemical correlation of an ash in the upper Malpais Tml section with the Ibapah Badlands type section (whose source was the southwestern Nevada volcanic field, Perkins and others, 1998) at 13.1±0.1 Ma. The Malpais Tml section is capped by the 11.9–8.8 Ma basalts that cap the Flattop area (Faulds and others, 2001b).

The tephra-correlation age suggests the Willow Beach beds are younger than Williams' Tml unit and represent axial-basin deposits, similar to the Lost Cabin beds, filling a tectonic sag or topographic low east of Willow Beach. Additional work is needed in correlating these deposits with other pre-Colorado River deposits. More analysis of tephra samples is planned to better constrain the age of the Willow Beach beds.

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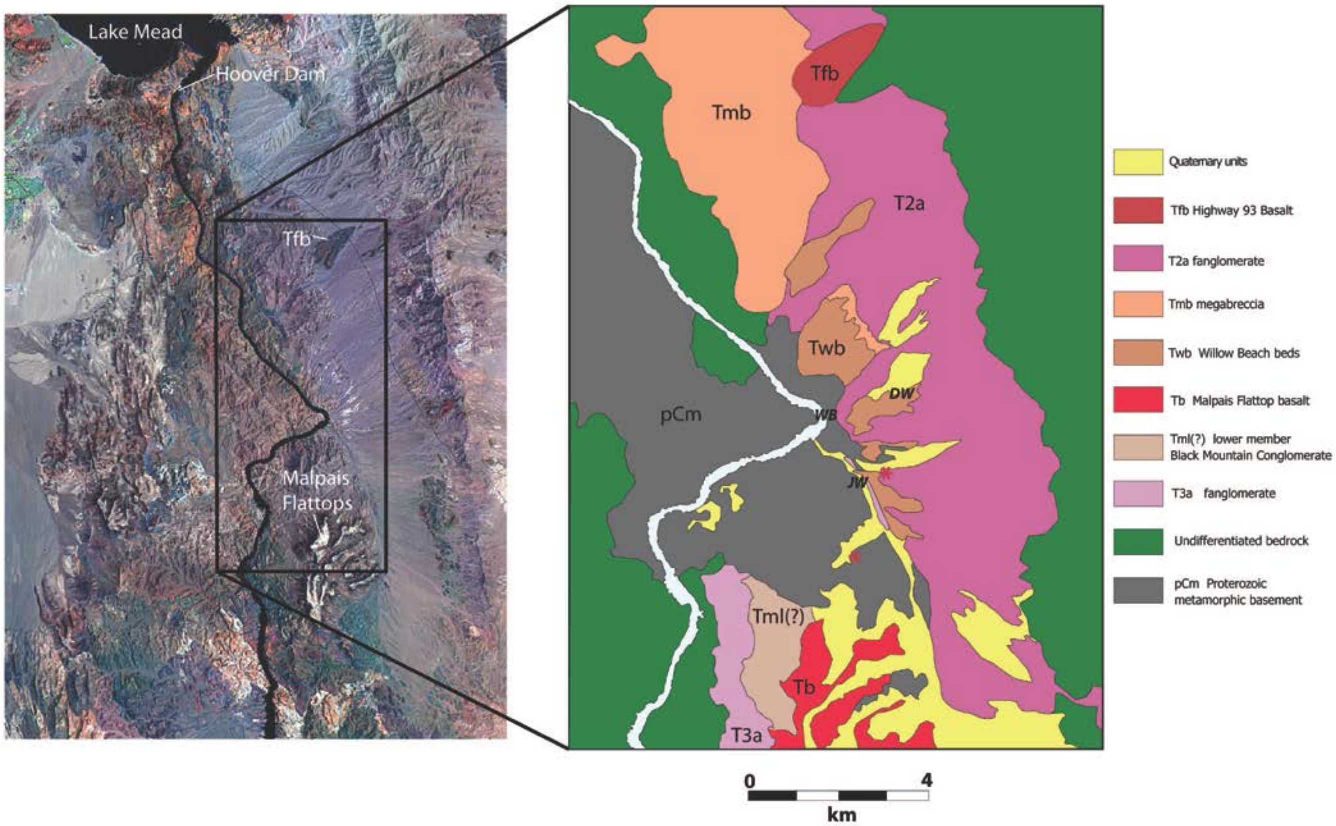

Figure 1. Location and geologic map of the Willow Beach area south of Lake Mead, the left photograph is Landsat imagery of Nevada and Arizona. The study area in the Black Mountain Basin lies within the Colorado River extensional corridor. Right is a generalized portion of the Boulder City 1:100,000 surficial geologic map showing the extent of the Willow Beach beds (Twb) that lie between the two areas described in Williams' (2003) thesis. Small exposures of late Mio-Pliocene Colorado river gravels are marked by a red asterisk. Other geographic features mentioned in the text are Willow Beach (*WB*), Jumbo Wash (*JW*), Disposal Wash (*DW*).

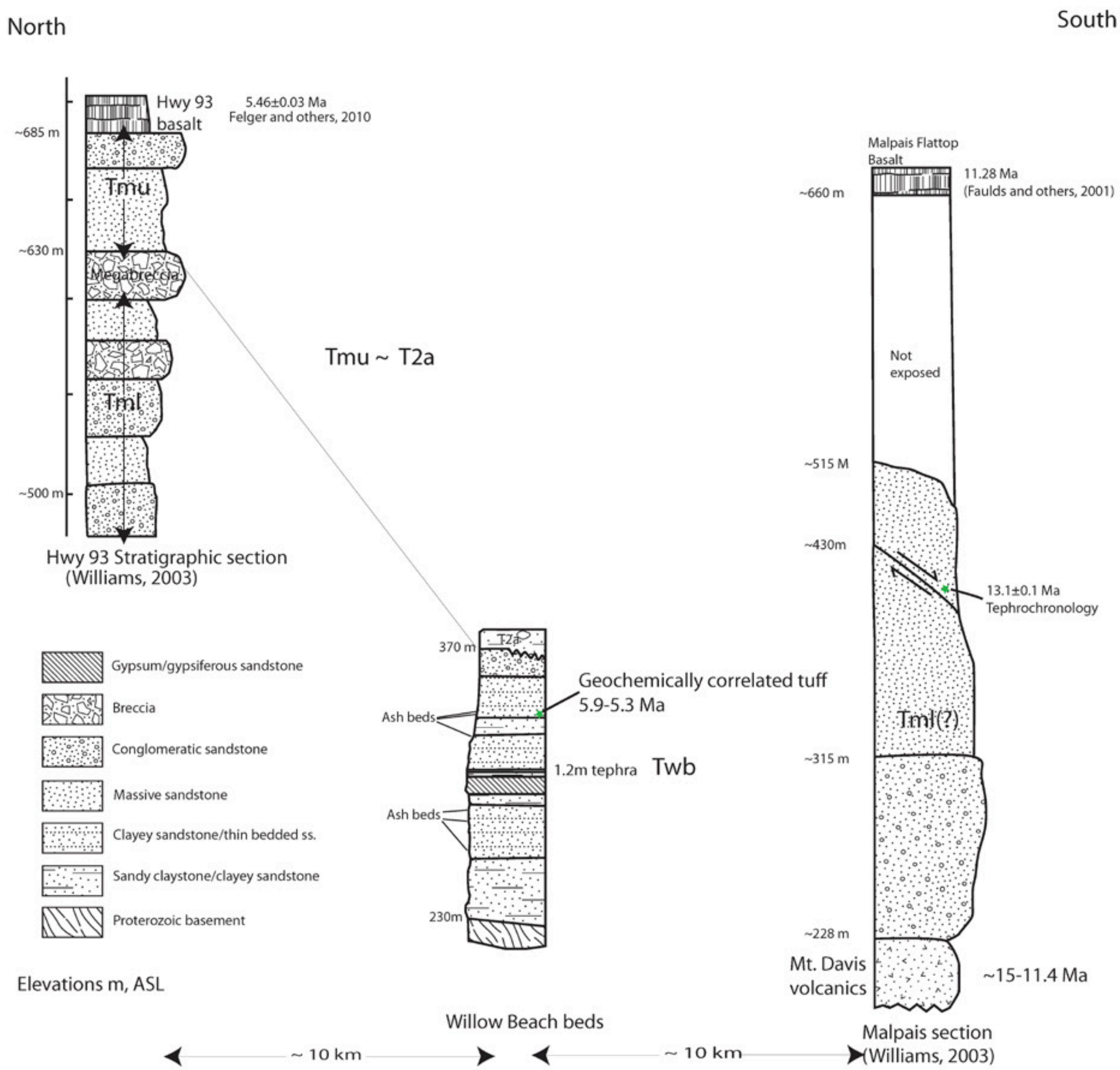

Figure 2. Measured stratigraphic section of Tmu and Tml measured along Hwy 93 (left) and Tml(?) on the west side of the Malpais Flattops (right) by Williams (2003), compared with the Disposal Wash (Twb) measured section (center). Tml in the Northern Black Canyon assemblage section further to the north is more than 400 m thick in a downfaulted part of the basin. The Malpais section mapped as Tml(?) is about 400 m thick, not accounting for the faulted section. There are significant differences in lithology, however, between Twb and Tml from both sections measured by Williams. The Twb deposits also are lower in elevation than the two Tml sections and may represent the deepest part of the extensional basin or may be downfaulted (elevation above sea level in brackets). The Disposal Wash measured section rests unconformably on T3a fanglomerate and is about 140 m thick. T2a appears correlative with Tmu from Williams (2003) mapping. A preliminary date of 5.9 to 5.3 Ma from geochemical correlation on the uppermost ash in the Twb section suggests the Willow Beach beds are younger than the nearby Tml deposits.

Origin of the Ancestral Colorado and Gunnison Rivers and Post-10 Ma River Incision Rates in Western Colorado

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Previous workers have advocated a late Miocene age for the origin of the upper Colorado River (Izett, 1975; Larson and others, 1975; Kirkham and others , 2002; Kunk and others , 2002; and Buffler, 2003). This report 1) summarizes evidence for late Miocene Colorado and Gunnison Rivers in the upper Colorado basin, and 2) discusses how Neogene uplift has influenced patterns of post-10 Ma river incision in western Colorado.

Ancestral Colorado River

River gravels associated with a ca. 10 Ma Colorado River are present at three important locations in western Colorado: 1) Wolford Mt (Kremmling), 2) Lookout Mt (Glenwood Canyon), and 3) Grand Mesa (fig. 1). At Wolford Mt Izett (1975) reported probable Colorado River gravels consisting of rounded Precambrian clasts that overlie an 11.2 Ma volcanic ash (apatite-fission-track age) in the Miocene Troublesome Formation. Near Glenwood Canyon, Larson and others (1975) reported ancient Colorado River gravels associated with 10 Ma basalt flows on Lookout Mountain. The gravels consist primarily of resistant quartzite but contain a significant quantity of granitic clasts that were probably derived from the nearby Sawatch or Gore Ranges (minimum transport distance of ~50 km). These deposits have been affected by post-10 Ma evaporite tectonism and are present at an elevation (elev. 2600 m) that is probably several hundred meters lower than their original position (Kirkham and others, 2002). Subsequent basaltic magmatism ca. 8-7 Ma is thought to have blocked the path of the ancestral Colorado River and diverted it northward to its present-day position in Glenwood Canyon (Kirkham and others , 2001).

Newly recognized river gravels (elev. 2935 m) located beneath radiometrically dated basalt flows of Grand Mesa confirm the presence of the ancestral Colorado River in western Colorado by ca. 11 Ma (Czapla and Aslan, 2009) (fig. 1). The river gravels are represented by 3 to 6 m of rounded pebbles and cobbles consisting of abundant quartzite, well cemented sandstone, and significant (~5%) quantities of granite. The nearest granitic outcrops at comparable elevations are located a minimum of ~150 km to the east in the Sawatch, Park, and Gore Ranges. The basal flow overlying the river gravels is dated to 10.76 Ma (Kunk and others, 2002). Collectively, these deposits strongly suggest that the ancestral Colorado River flowed west from the Rockies out onto the Colorado Plateau by ca. 11 Ma.

Ancestral Gunnison River

River gravels associated with a ca. 10 Ma Gunnison River in western Colorado are present at two locations: 1) Flat Top Mountain and 2) the Uncompahgre Plateau (fig. 1). At Flat Top Mountain near Gunnison, CO, ancestral Gunnison River gravels consist of ~15 m of rounded sandy to cobble-rich gravel that underlies ca. 10 Ma basalt flows (A. Stork and CREST unpublished data). The lithologies of the high-elevation (3078 m) gravels are dominated by Precambrian rock types similar to those in the nearby Sawatch Range located <50 km east of Flat Top Mountain.

Ancient river gravels representing deposits of the ancestral Gunnison and Uncompahgre Rivers are present on the Uncompahgre Plateau (Betton and others, 2005; Aslan and others, 2008a). The oldest river gravels are represented by rounded pebble to cobble gravel remnants scattered along the crest of the northwest-trending Uncompahgre Plateau at elevations ranging from 2500 to 3000 m (fig. 1). Gravels are found as far north as the southern edge of Unaweep Canyon. These gravels are dominated by intermediate volcanic clasts representing Tertiary rocks of the San Juan and West Elk Mountains. North of Goddard Creek, however, there are small but significant percentages of granitic clasts in the river gravels. There is no obvious source for these granitic rocks in the Uncompahgre River basin or nearby San Juan Mountains. There are, however, abundant sources of granite in the Gunnison River basin to the east. Based on these observations, it is plausible that these granite-bearing river gravels represent deposits of an ancestral Gunnison River, which flowed west from present-day Gunnison, CO to the Uncompahgre Plateau. The ancestral Uncompahgre River probably joined the ancestral Gunnison River near Goddard Creek, and the combined rivers then flowed northwest to Unaweep Canyon where they joined the ancestral Colorado River (fig. 1). Similarities in the elevations of the ancestral Colorado River gravels on Grand Mesa, the ancestral Gunnison-Uncompahgre River gravels on the Uncompahgre Plateau, and the bedrock rim of Unaweep Canyon suggest that these ancient river systems collectively flowed west onto the Colorado Plateau ca. 11 Ma. Abandonment of Unaweep Canyon occurred ca. 1 Ma (Aslan and others, 2008b).

Post-10 Ma River Incision Rates

Widespread ca. 10 Ma basalt flows and selected volcanic ashes serve as a datum for calculating post-10 Ma river incision rates in western Colorado (fig. 2). The main picture that emerges from the river incision data is one of regional variability, which we hypothesize to reflect differential uplift of the Colorado Rockies since the late Miocene. Three broad areas representing different rates and magnitudes of post-10 Ma river incision are identified: 1) Grand Mesa and the Flat Tops, 2) the Elkhead Mountains and northern Park Range, and 3) Browns Park and the Sand Wash Basin.

Maximum rates (100-150 m/my) and magnitudes (1000-1500 m) of river incision are recorded between Grand Mesa and Glenwood Canyon, and in the Flat Tops north of Glenwood Canyon. Incision rates and magnitudes decrease (rates <100 m/my and magnitudes range from 840 to 290 m) to the north in the vicinity of the Elkhead Mountains and the northern Park Range. Minimum rates (<50 m/my) and magnitudes (<250 m) of river incision are associated with Browns Park and the Sand Wash Basin in northwestern Colorado. Slow incision rates and low incision magnitudes also occur in Middle Park (fig. 2).

The observed variability of river incision rates and magnitudes argues against climate as the primary driver for post-10 Ma incision. If climate were the primary driver, one would expect more similar incision values given the proximity of the localities relative to one another. Also, similarity of long- and short-term river incision rates in selected areas argues for semi-steady river incision due to base level lowering (Aslan and others, 2008a). We see no strong evidence of accelerated incision rates that coincides with the onset of glacial climates ca. 3-4 Ma.

Instead, the wide variation in rates and magnitudes of river incision are consistent with differential uplift of subregional blocks of the Colorado Rockies during the late Cenozoic. These structural influences can be inferred by comparing incision rates and magnitudes at locations upstream and downstream of important knickzones (fig. 2). For example, along the Colorado River, post-10 Ma rates and magnitudes of incision remain fairly constant (rates >100 m/my; magnitudes >1000 m) from Grand Mesa upstream to Gore Canyon, and then decrease markedly (rates <10 m/my; magnitudes <100 m) across the Gore Canyon knickzone in Middle Park. This pattern seems to reflect an upstream wave of incision caused by regional uplift and/or downstream base-level fall. We favor an uplift mechanism because ca. 10 Ma deposits of the Troublesome Formation in Middle Park are part of a down-faulted block, and normal faults along the eastern side of the Gore Range show post-Laramide movement. In addition, river incision in this region began ca. 10-8 Ma, prior to integration of the Colorado River system through Grand Canyon.

In contrast to the Colorado River example, river incision rates and magnitudes are low (rates 15-27 m/my; magnitudes < 230 m) immediately upstream of Yampa Canyon knickzone, and then increase significantly (rates 132-180 m/my; magnitudes ~1700 m) upstream near the headwaters (fig. 2). We interpret this upstream increase in river incision rate and magnitude to reflect post-Miocene uplift of the Yampa River headwaters (i.e., the Flat Tops) relative to the Browns Park-Sand Wash Basin regions.

Conclusions

1. The ancestral Colorado and Gunnison Rivers were flowing west from the Colorado Rockies onto the Colorado Plateau by ca. 11 Ma.

2. Post-10 Ma river incision rates and magnitudes in western Colorado show a wide range of variability reflecting post-Laramide differential uplift. Fast incision rates (100-150 m/my) and large magnitudes of incision (1.0-1.5 km) are associated with basalt plateaus such as Grand Mesa and the Flat Tops. It is plausible that these areas have undergone 1.0-1.5 km of rock uplift since 10 Ma, which possibly translates to ~1.0-1.5 km of surface uplift because Grand Mesa and the Flat Tops have resistant basalt caps. Regions such as Middle Park, Browns Park, and the Sand Wash Basin have experienced slow rates (<50 m/my) and small magnitudes (<250 m) of post-10 Ma river incision. During late Cenozoic regional uplift, these areas were less elevated relative to the surrounding mountain ranges due to normal faulting along their perimeters.

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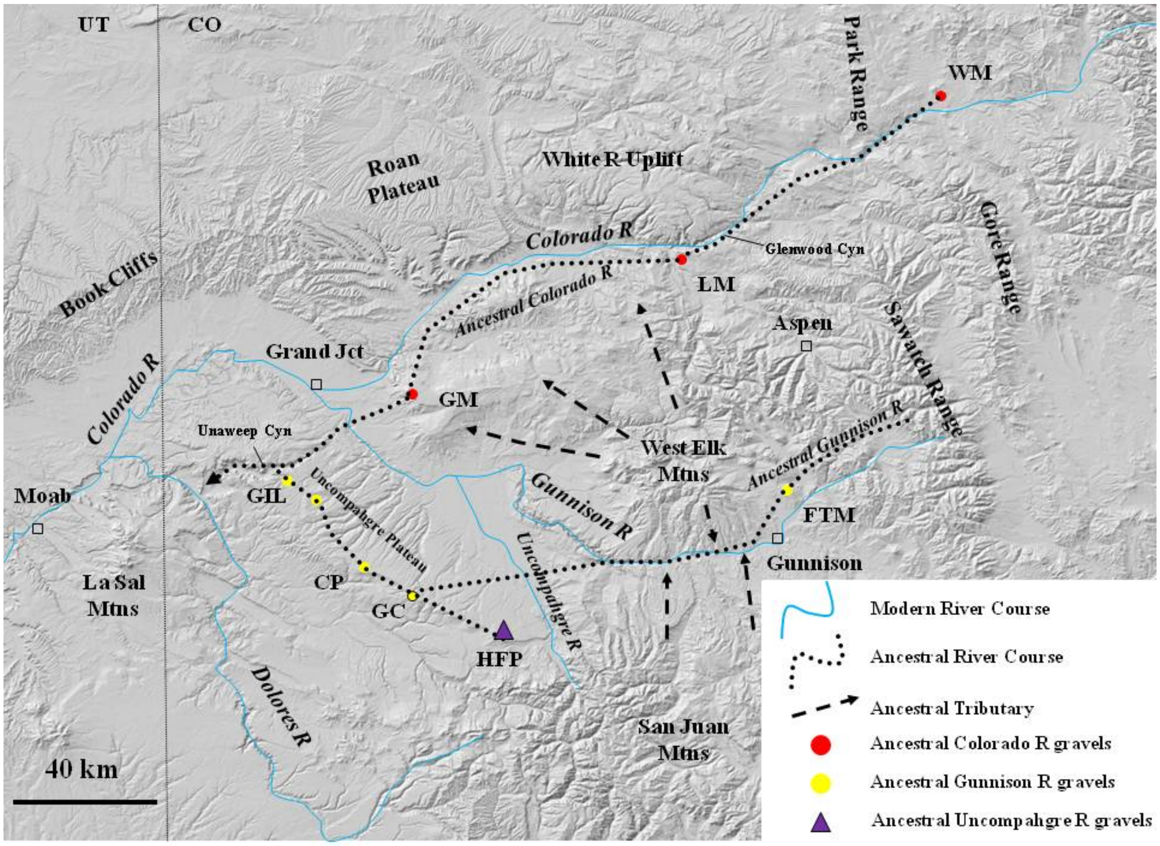

Figure 1. DEM showing important locations of high-elevation river gravel deposits and the ca. 11 Ma courses of the ancestral Colorado and Gunnison Rivers in western Colorado. Data compiled from Izett (1975), Larson and others (1975), Betton and others (2005), Aslan and others (2008a), Czapla and Aslan (2009), and unpublished CREST data. CP, Columbine Pass; FTM, Flat Top Mt; GC, Goddard Creek; GIL, Gill Creek; HFP, Horsefly Peak; LM, Lookout Mt; WM, Wolford Mt.

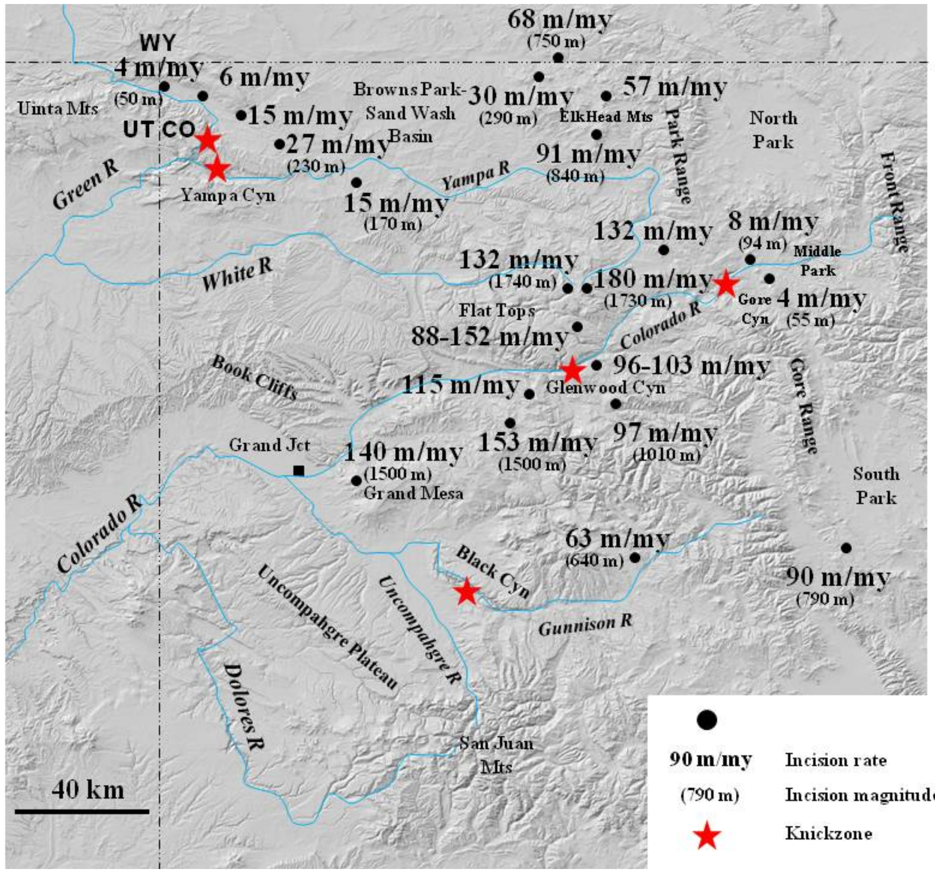

Figure 2. DEM showing rates and magnitudes of river incision in western Colorado averaged over the past ca. 10 Ma. Data are calculated using age estimates of basaltic flows and volcanic ashes that are associated with Miocene fluvial deposits. Basalt flows range in age from 6 to 14 Ma although the majority of the flows are 10 Ma. Data are from Izett (1975), Larson and others (1975), Luft (1985), Kunk and others (2002), Buffler (2003), Aslan and others (2008a), and unpublished data.

Kingman Uplift, Paleovalleys and Extensional Foundering in Northwest Arizona

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Kingman Uplift

The Kingman uplift, northwest Arizona and southeast Nevada, is a large latest Cretaceous and/or early Tertiary (Laramide) structural high along the southwest edge of the Colorado Plateau. The uplift was named by Lucchitta in Goetz and others, 1975, and it is well documented and described by Young, 1979, 1985, 2001; Lucchitta and Young, 1986; Bohannon,1984; Beard, 1996; and Faulds and others, 2001. It lies between the Colorado Plateau and east of the Cretaceous-age Sevier thrust belt (fig. 1). The uplift is defined by Tertiary volcanic and sedimentary rocks directly overlying Proterozoic and Cretaceous crystalline basement rocks; Paleozoic and Mesozoic rocks have been eroded from the uplift. The Hualapai Plateau part of the Colorado Plateau on the east side of the uplift exhibits a progressive westward beveling of the middle and lower Paleozoic section, which formed a smooth surface with slight relief developed along the truncated edges of the resistant Muav and Redwall Limestones. Paleocanyons cut into the Hualapai Plateau contain conglomerates with clasts and unroofing assemblages indicating derivation from the Kingman uplift to the west (Young, 2001). Erosional retreat of the Upper Permian Kaibab-Toroweap Formations, in contrast to the subdued beveled topography of the middle and lower Paleozoic rocks, formed a southwest facing scarp that is still visible today on the north side of the Grand Canyon. Young (1985) showed that the Kaibab-Toroweap scarp has retreated no more than ~8 km to the northeast since ~19 Ma, to its present position on the north side of the Grand Canyon.

The north end of the Kingman uplift is a north-plunging arch with similar southward beveling of Paleozoic strata, preserved below pre-extension Tertiary strata (Bohannon, 1984, Lucchitta and Young, 1986). Beard (1996) suggested that the Permian Kaibab-Toroweap scarp continued along the north-plunging limb of the arch and formed a topographic barrier between the Proterozoic and lower Paleozoic rocks on the Kingman uplift and lower topography to the north of the scarp. The west margin of the uplift is somewhere east of the Spring Mountains, where the autochthonous Paleozoic section below the Sevier thrusts is beveled progressively to the southeast. Herrington (1993) described east to northeast flowing paleochannels cut into Proterozoic rock southeast of Sheep Mountain (fig. 1), containing locally-derived conglomerates as much as 100 m thick and locally capped by ~18 Ma volcanic rocks.

The Kingman uplift is most likely a Laramide-age basement-cored uplift. Faulds and others (2001) suggested that the northern Grand Wash and Cerbat Mountain faults on the northeast side of the uplift could be reactivated, west-dipping, Laramide reverse faults (fig. 2). The Kingman uplift is younger than 64 – 73 Ma plutons in the Black Mountains (Faulds and others, 2001) and older than, or synchronous with, formation of paleocanyons that contain sediments as old as Paleocene (Young, 2001) along the Plateau margin. Fitzgerald and others (2009) interpreted the onset of Laramide cooling as ~75 Ma. On the west flank of the uplift, Herrington (2000) used fission track and $^{40}Ar/^{39}Ar$ thermochronology to suggest removal of the Mesozoic section by ~85 Ma, subsequent stripping of the Paleozoic section by 40 Ma, and cutting of the paleochannels across the uplift before 18 Ma. Faulds and others (2001) indicated that two-mica, garnet-bearing 64 – 73 Ma plutons in the Black Mountains were emplaced at depths as great as 10 km; this contrasts with the shallow nature of contemporaneous plutons in the Cerbat and Hualapai Mountains and along the Grand Wash Cliffs. For example, the 65.5 ± 3.5 Ma Clay Spring pluton intrudes the base of the Paleozoic section and probably was emplaced no more than at a few kilometers (Young, 2001) because apatite fission-track cooling ages as old as 90 to 110 Ma, sampled just below the basal Paleozoic unconformity, indicate residence in a partial-annealing zone prior to onset of rapid cooling at about 70 – 50 Ma (Fitzgerald and others (2009).

Volcanic rocks ~18 Ma or older, deposited across the beveled edge of the lower to middle Paleozoic rocks on the Hualapai plateau margin, indicate that: (1) most Mesozoic and late Paleozoic strata were stripped from the Plateau margin by 18 Ma, and (2) the beveled surfaces have not retreated significantly since 18 Ma, except immediately adjacent to the Colorado River and its tributaries and, therefore, are subject to erosion related to downcutting of the Colorado River. Distribution of volcanic rocks from sources west of the plateau, such as the ~19 Ma Separation and ~17.4 Ma Iron Mountain basalts and 18.5 Ma Peach Spring Tuff, indicate that the Cerbat and Hualapai Mountains were still high relative to the plateau at ~17 – 18 Ma (fig. 3). Because truncation predates extension, the erosional edges of the Plateau can be projected into the extended terrane of the Lake Mead region. Figure 1 refines the reconstruction of the Kingman uplift by using well-documented truncations and following Young (1979, 1985, 2001), Bohannon (1984), and Beard (1996). The truncation pattern on the southwest margin of the plateau, from Iron Mountain to the Shivwits plateau, is used to project the feather edge of the Paleozoic section around the nose of the uplift; specifically, we use the distance, about 10 km, between the southwest edges of Mississippian Redwall Limestone and Cambrian Tapeats Sandstone. At Sheep Mountain, on the west side of the uplift, the Proterozoic rocks are overlain by Cambrian through Mississippian rocks (Herrington, 2000); the estimated 10 km distance would put the edge of the Tapeats on the west flank of the McCullough Mountains at ~20 Ma. Projections at the south end of the River Mountains and on Wilson Ridge (fig. 1) are based on the presence there of a thin section of altered and intruded Cambrian through Mississippian rocks. Therefore, the southern limit of the Tapeats Sandstone is most likely a minimum.

Paleocanyons

From south to north, the Peach Springs, Milkweed, and Meriwitica paleocanyons are cut into the Hualapai Plateau on the east flank of the Kingman uplift (fig. 3). These canyons contain Paleocene and younger sedimentary fill capped by volcanic rocks, including the 18.5 Ma Peach Spring Tuff, an ash-flow tuff erupted from the Silver Creek Caldera in the southern Black Mountains (fig. 3; Ferguson, 2008). The Peach Springs paleocanyon is the deepest and broadest, ~1200 m deep, and 5 km wide at the bottom and widening upward into a valley about 15 km wide at the rim (Young, 2001). At Truxton Canyon on the edge of the plateau, the paleocanyon is rimmed by Tapeats Sandstone, and northeast at Peach Springs, it is rimmed by Muav Limestone. This canyon and Milkweed, the next paleocanyon north, include sediments at the bottom derived from the Kingman uplift to the west, unconformably overlain by locally derived conglomerate (Young, 1989). The Milkweed paleocanyon is only a few kilometers wide at the rim and includes about 360 m of Paleocene and younger sediment (Young, 2001). The Meriwitica paleocanyon is shallow, and only thin local sections of sediment are preserved beneath the Peach Spring Tuff and the 19.03 Ma Separation basalt. The basalt originated from west of the Plateau margin and flowed northeast, where the furthest outcrop lies approximately 13 km from the Kaibab-Toroweap cliff (fig. 1) and predates the cutting of Spencer Canyon, a major tributary to the Colorado River.

A broad east-northeast-trending paleovalley lies between the Cerbat and Hualapai Mountains at Kingman. The floor of the paleovalley is broken by scattered basement highs surrounded by up to 200 m of locally and more distally derived Cenozoic mafic volcanic flows (Buesch and Valentine, 1986). The volcanic rocks are overlain by as much as 100 m of Peach Spring Tuff, which flowed from west to east through the paleovalley from its source in the Black Mountains, lapping against the crystalline outcrops of the Cerbat and Hualapai Mountains. The south edge of the paleovalley is fairly steep against the Hualapai Mountains, where total paleorelief is 1,200 m or more. In contrast, the north edge slopes gently up the flank of the Cerbat Mountains, with about 700 – 800 m of paleorelief to the crest of the range. Volcanic rocks in the paleochannel and on the east side of the Cerbat Mountains dip gently eastward into Hualapai Valley. North of the paleovalley, Peach Spring Tuff occurs in three isolated outcrops on the east side of the Cerbat Mountains. East of Kingman, a northeast-trending valley and volcanic rocks in the subsurface suggest the paleovalley continued eastward towards the Hualapai Plateau.

Surface profiles of this paleovalley and the Peach Springs paleovalley on Hualapai Plateau are strikingly similar in width and depth, based on the configuration of the top of the Peach Spring Tuff, suggesting the two paleovalleys represent the same system (fig. 4). Young and Brennan (1974) assumed that the Peach Spring Tuff was not channelized and fanned out across a gently east-sloping terrain, characterized by broad valleys of low relief between the source and the Hualapai Plateau. However, correlation of paleovalleys suggests an alternative interpretation. North of the paleovalley, Peach Spring Tuff is found only in local outcrops on the east side of the Cerbat Mountains. We suggest that the north side of the Kingman paleovalley extends southwest at least to the southern tip of the Black Mountains, and northeast to at least Long Mountain (fig. 3). If so, the two northern outcrops of Peach Spring Tuff along the east side of the Cerbat Mountains could mark paleocanyons that correspond to the Milkweed and Meriwitica valleys on the Hualapai Plateau. Speculatively, the broader exposures of Peach Spring Tuff at the south end of the Black Mountains may track the southwest continuation of the Kingman paleovalley. Lack of sediment beneath the volcanic rocks in the Kingman paleovalley indicates any Paleocene deposits were removed prior to about 20 Ma, the age of the volcanic rocks.

More problematic are the Peach Spring Tuff outcrops along Interstate 40 and on the west flank of the Mohon Mountains, south of the Peach Spring paleovalley. The Peacock Range lies between the Kingman to Peach Spring paleovalley. West-side-down faults on the west side of the Peacock Range (fig. 1) accommodate downthrow and east-tilting of basalt flows and the Peach Spring Tuff. The Peacock Range itself is a fault block, down-dropped and east-tilted along the southern Grand Wash fault (fig. 4). Therefore, uplift of the Peacock Range postdates the paleocanyon. Volcanic rocks deposited around the south end of the range, including the Peach Spring Tuff, suggest generally low relief between Long Mountain and the north end of the

Hualapai Mountains, as shown by Young and Brennan (1974), to allow the Peach Spring Tuff to flow into the Peach Spring paleovalley and along the I-40 corridor (figs. 1 and 3).

Foundering of Kingman Uplift

We suggest that the highest part of the Kingman uplift foundered and evolved into the northern Colorado River Extensional corridor, as large-magnitude extension widened the central part of the preexisting highlands beginning ~16 Ma (fig. 5). The eastward limit of synextensional volcanism is marked by the South Virgin-White Hills detachment fault and by a poorly defined boundary west of the Cerbat and Hualapai Mountains. On the Hualapai Plateau, Young (2001, this volume) documented an abrupt shift from sediments derived from south and west of the plateau to locally derived conglomerates in early Miocene time, prior to the onset of major extension. That shift could be related to either early mild extension or to significant erosional degradation of the highlands, such that stream gradients were greatly reduced. Either way, middle Miocene extension greatly disrupted the topography and ultimately facilitated a drainage reversal and development of the southwest-flowing Colorado River drainage along the north side and then through the center of the Kingman uplift and northern Colorado River extensional corridor.

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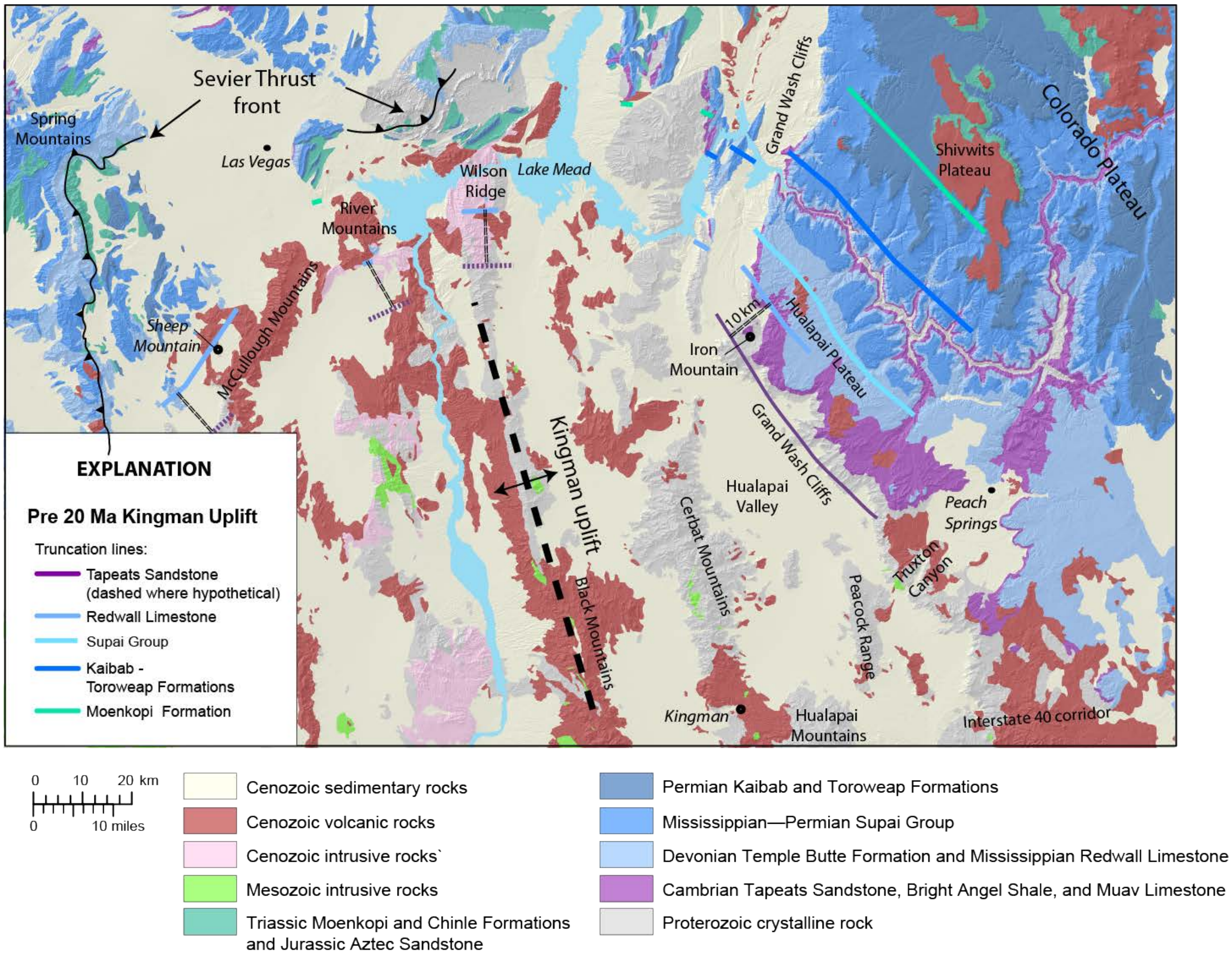

Figure 1. Geologic map of northwest Arizona and southeast Nevada, showing reconstruction of the arch around north end of Kingman uplift. Reconstruction is based on the southwest truncations of Paleozoic strata on the southwest margin of the Colorado Plateau, which are then projected into the Basin and Range. Horizontal distance between Tapeats Sandstone and Redwall Limestone on Hualapai Plateau of about 10 km is used to project southward truncation of Tapeats around the north end of the arch. Generalized geology is overlain on 30 m DEM.

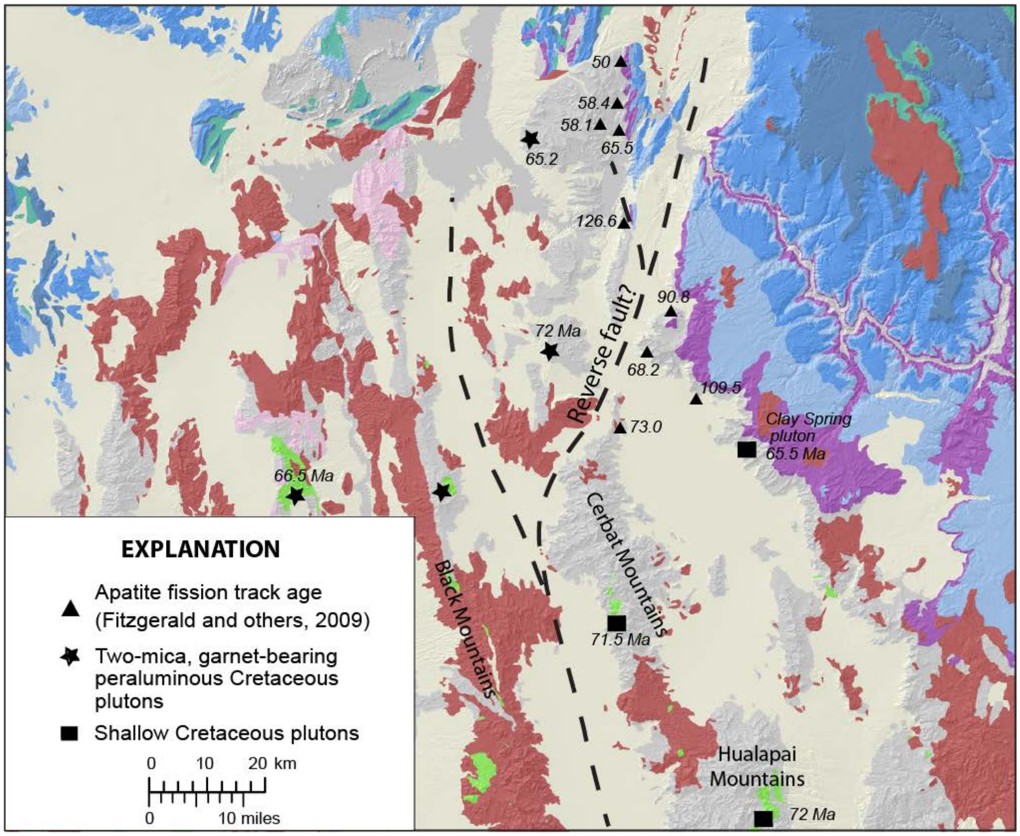

Figure 2. Geologic map (same as fig. 1), showing possible Laramide or older reverse fault (or faults) as proposed by Faulds and others (2001). Note the 126.6 Ma apatite fission-track age (Fitzgerald and others, 2009) to west of the labeled fault, compared to the 90.8 Ma age to east. Both samples were collected withing 200 m of the base of the Paleozoic section.

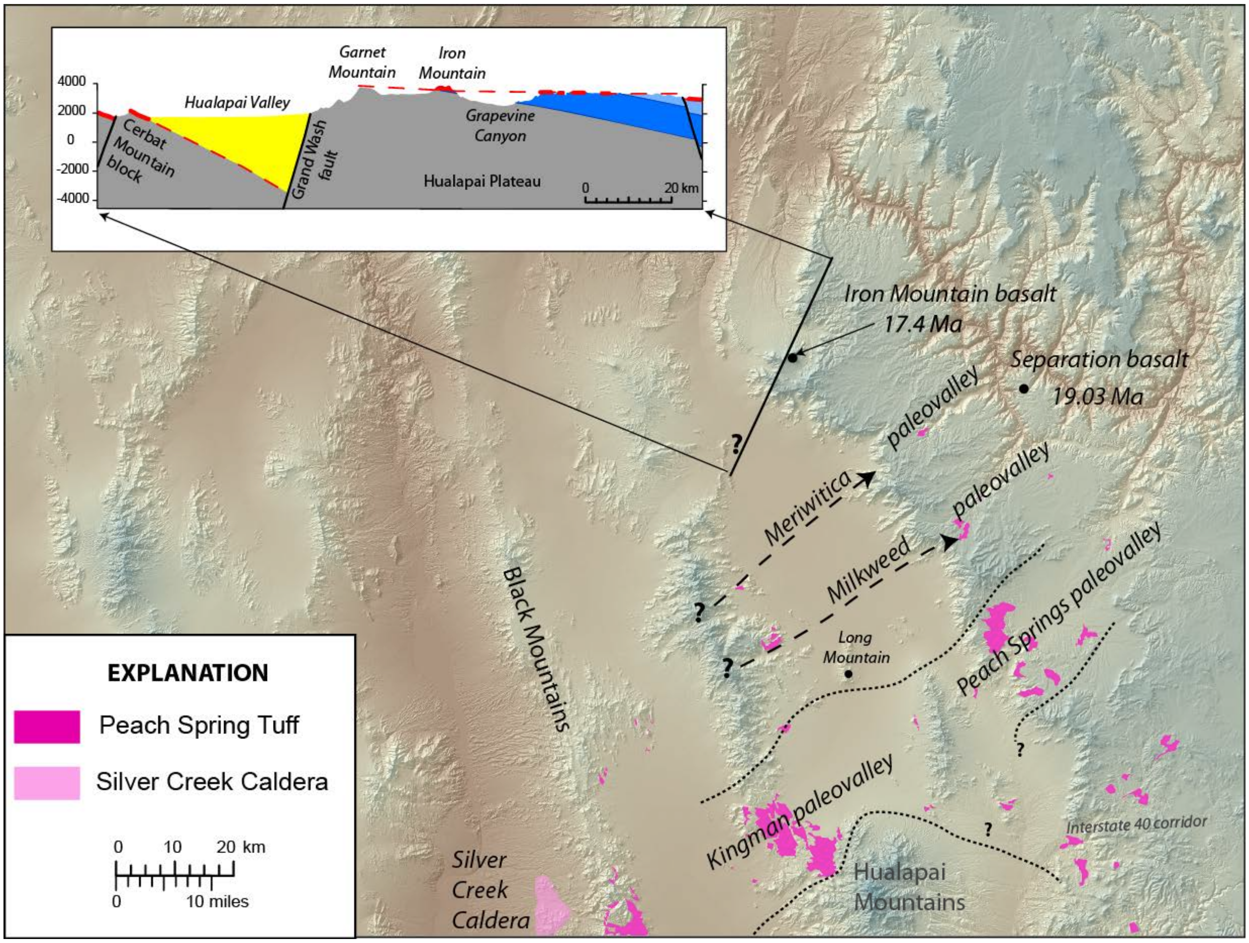

Figure 3. Shaded-relief map of northwest Arizona and southeast Nevada, showing Kingman and Peach Springs plaeovalleys and location of 18.5 Ma Peach Spring Tuff (Ferguson, 2008). Westward projection of Milkweed and Meriwitica paleovalleys is speculative. Profile of Iron Mountain flow indicates it flowed across Grapevine Canyon and is cut by the Grand Wash fault. Similarly, Separation flow predates cutting of Grand Canyon tributaries (Lucchitta and Young, 1986).

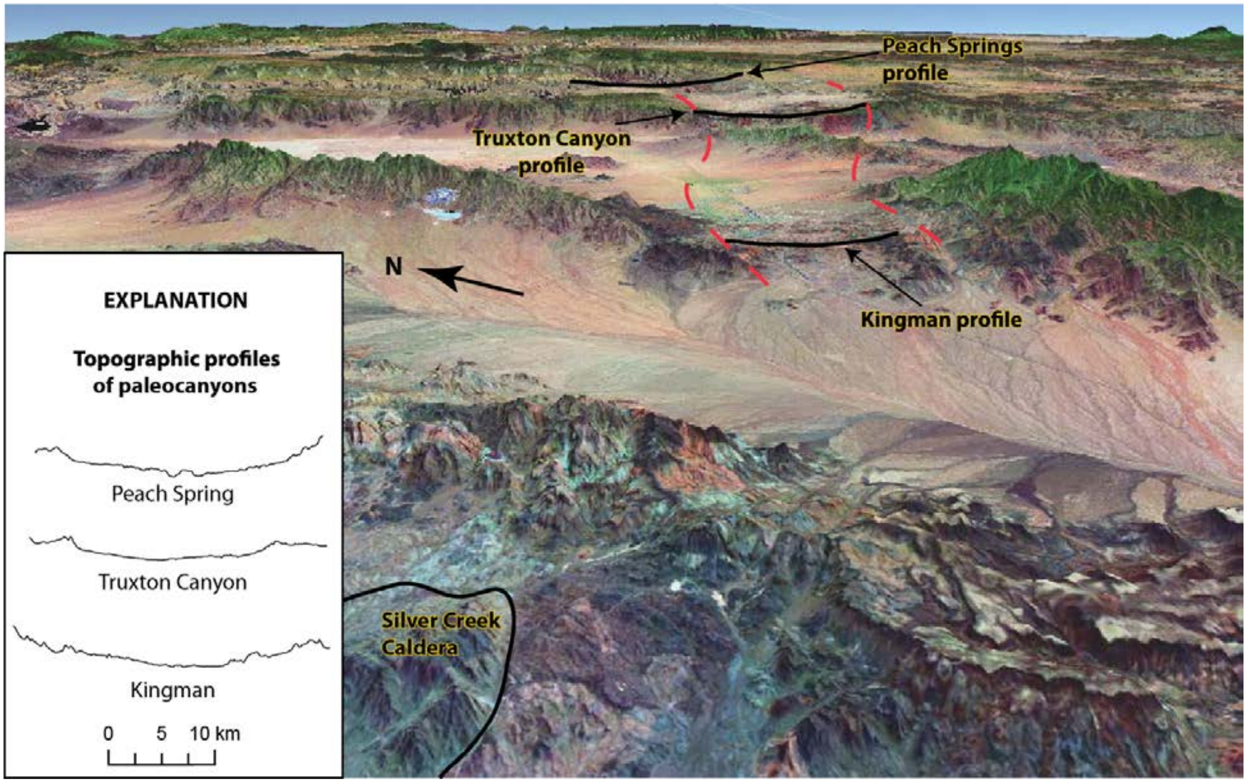

Figure 4. View to the east from Black Mountains across Kingman paleovalley to Hualapai Plateau. Cerbat Mountains are to the left of Kingman, the Hualapai Mountains are to the right. Silver Creek Caldera is the inferred source of Peach Spring Tuff. Perspective view is from NASA Worldwind (*http://worldwind.arc.nasa.gov/java/*).

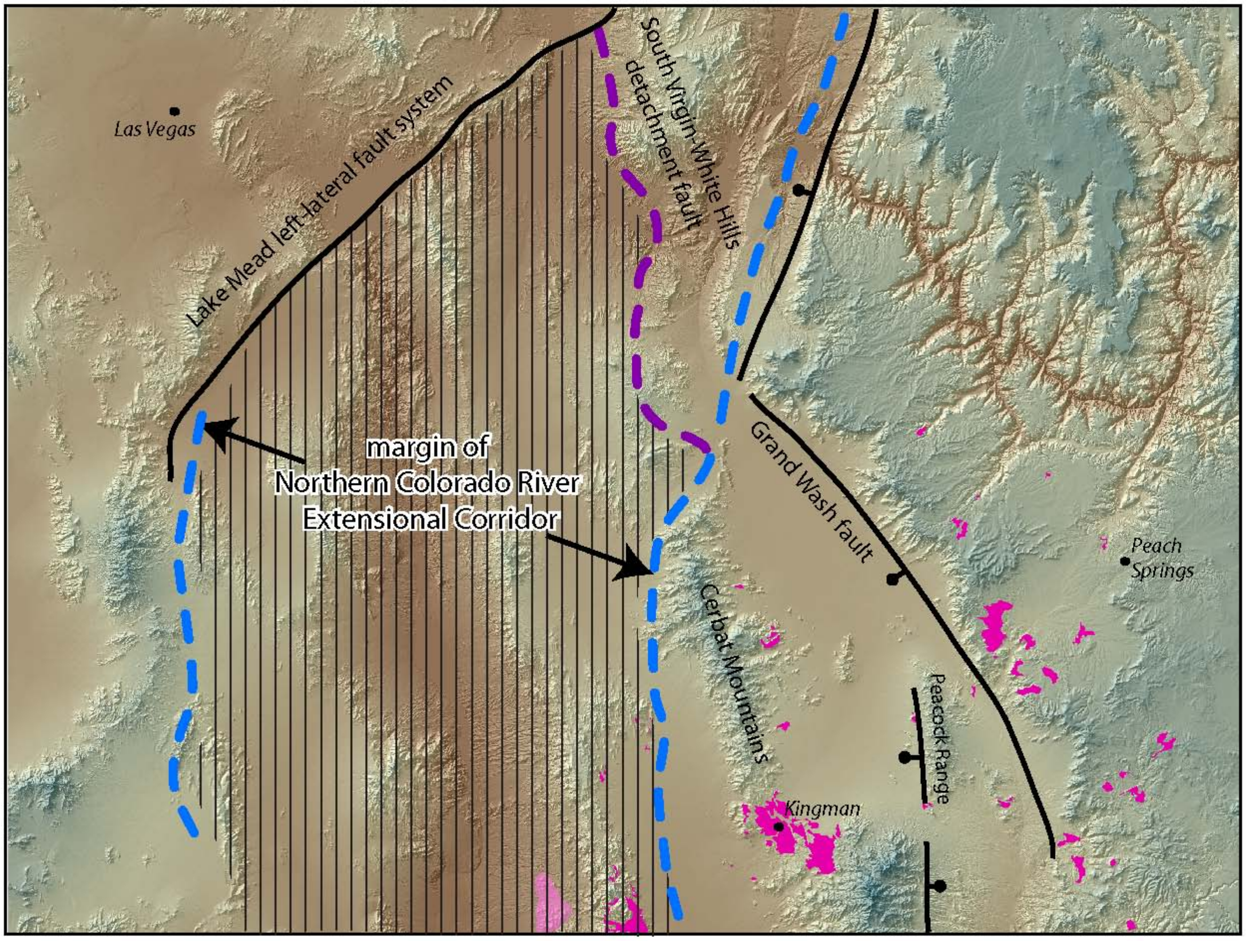

Figure 5. Shaded-relief map showing the location of the northern Colorado River extensional corridor (blue dashed line; modified from Howard and John, 1987) and the eastern limit of extensive volcanism (purple dashed line). Pink areas are the same as in fig. 3.

Oligocene-Early Miocene Incision, Strike-Valley Development, and Aggradation, Mogollon Rim, Verde Valley Region, Arizona—A Potential Analogue for pre-Grand Canyon Development

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Pre-Miocene incision occurred in the Verde Valley region across the present Mogollon Rim prior to 26.4 Ma. Regional geomorphic, structural, and stratigraphic relations suggest the presence of the Mogollon Rim by this time (Late Oligocene) as well as deep stream incision and southward-directed drainage along the margin of the Colorado Plateau. These relations, first synthesized and published by Peirce and others (1979), have been expanded using new stratigraphic and sedimentologic data. The best exposures and most complete stratigraphic sections that illustrate these points are located at the foot of the Mogollon Rim along AZ State Highway 179 and I-17, Yavapai County, AZ. Especially notable is the so-called Beavertail Butte locality at the junction of AZ 179 and the Beaverhead Flats Road (fig. 1).

In the vicinity of Beavertail Butte, the boundary between the Paleozoic - Cenozoic is marked by prominent local relief of several hundred meters, by an obvious discordance or angularity that everywhere dips to the south (a buttress unconformity), and by an onlap relation of Cenozoic conglomerate onto underlying Permian redbeds. These geomorphic relations document considerable local relief and streamflow to the south, two critical points in unraveling Cenozoic history of the region (Loseke, 2004; Loseke and Blakey, 2001; Ranney, 1988). The Cenozoic sedimentary rocks have been assigned to the Beavertail Butte formation (see Loseke, 2004 for discussion of nomenclature for this informal stratigraphic unit). The oldest Cenozoic valley fill consists of pale yellowish-tan, well-bedded conglomerate and local sandstone (Lookout Mountain member). Clasts locally range to large boulder size (> 1m dia.) but mainly comprise rounded pebbles and cobbles; all clasts are exclusively derived from the adjacent Mogollon Rim and include primarily Kaibab Formation chert and dolomite and Coconino Sandstone clasts with minor sandstone clasts from the Schnebly Hill Formation. Both grain-supported (streamflow deposits) and non-grain-supported (debris flow deposits) fabrics occur in beds up to several decimeters thick. Total maximum thickness of the member is 30 m. The Lookout Mountain member grades both laterally and vertically into two mostly younger units.

At Beavertail Butte, the medial part of the Beavertail Butte formation consists of yellowish-tan to pale gray mudstone, limestone, marl, volcanic ash, and sandstone and forms the Winter Cabin member. Maximum measured thickness is 20 m, although local geomorphic relations suggest that the unit may be locally much thicker. Bedding is mostly obscured but ranges from crinkly to contorted laminations in local arroyo cuts. The member marks a dramatic change in depositional style and is interpreted as lacustrine and related low-energy stream, pedogenic, and spring deposits. The member grades laterally and vertically into the other two members.

The youngest member of the Beavertail Butte formation is the Little Pig member. It consists of mostly pale reddish-brown, well-bedded, pebble conglomerate with local cobble and boulder clasts to 1 m in diameter. Unit thickness is generally +/- 20 m. Clasts reflect a wide range of sources and lithology including Kaibab and Coconino clasts as in the Lookout Mountain member, older Paleozoic clasts from the Redwall, Martin, and Tapeats formations, Proterozoic clasts from the Precambrian of central Arizona (quartzite clasts are very rare), and distinctive latite clasts from the Oligocene Sullivan Buttes Latite. Virtually all identifiable clasts can be traced to sources currently present in central and northern Yavapai County and SW Coconino County. Clearly, a much broader range of clasts were available than was the case for the older Lookout Mountain member. The Little Pig member and similar conglomerates are presently distributed along the lower reaches of the Mogollon Rim from near Chino Valley to the NW to Fossil Creek to the SE and define a Rim-parallel strike valley incised mostly into the easily erodible Hermit Formation. This member is much more widely distributed than the other two members and at most other localities outside the corridor along Az179 to I-17 is the only member present. We interpret this distribution to mark the presence of a continuous, regional drainage system. The Little Pig member marks another dramatic shift in depositional style and is interpreted as a bedload gravelly stream deposit. The present outcrop trend marks a course similar to that of the modern Verde River.

The Beavertail Butte formation marks a major change in Cenozoic geologic history in central Arizona. Earlier Cenozoic rivers drained the Mogollon highlands to the south and flowed northward, presumably into southern Utah near Bryce Canyon. Some of these rivers probably deposited the so-called Rim gravels on the upper reaches of the Mogollon Rim (Potochnik, 2001). In contrast, the Beavertail Butte formation exclusively lies below the level of the Mogollon Rim (fig. 2) and contains clasts derived from it. Such a relationship strongly suggests the presence of the Rim by onset of Beavertail Butte deposition. It is possible that remnants of the Rim lay to the south as well and that these remnants were subsequently eroded and incorporated into the Beavertail Butte and related sedimentary units. All members of the Beavertail Butte formation as well as coeval conglomerates in the greater Verde Valley region are offset by extensional normal faults (Basin and Range faulting *ca.* 12 Ma – 2 Ma) that mostly trend NW-SE.

The age of the Beavertail Butte formation is moderately well constrained. The unit must postdate the youngest Rim gravels dated at *ca.* 33 Ma, Early Oligocene (Potochnik, 2001). A dated volcanic ash bed in the Winter Cabin member has yielded a multiple-crystal biotite $^{40}Ar/^{39}Ar$ age of 26.4 Ma (Late Oligocene). Undated clasts in the Little Pig member derived from the Sullivan Buttes Latite are *ca.* 26 Ma. Overlying basalts at Beavertail Butte have 15.4 and 13.2 Ma ages (Middle Miocene). Geomorphic evidence suggests that the regionally extensive Middle Miocene basalt episode (Hickey and House Mountain basalts) completely disrupted older Miocene drainage patterns. These patterns constrain ages of the Beavertail Butte formation between *ca.* 30 Ma and 15 Ma.

The evidence reported above strongly supports incision of paleovalleys and presence of a topographic Mogollon Rim in approximately its present position by *ca.* 26+ Ma. Initial fluvial/alluvial deposits drained the Mogollon Rim and flowed southward; the location of the mouths of these streams is unknown. Local to regional ponding formed lacustrine settings that were succeeded by a trunk strike valley stream that flowed from NW to SE (fig. 3). Most likely these streams drained into the Tonto Basin, a major Late Oligocene-Early Miocene depocenter in Central Arizona. The hogbacks or cuestas that bounded the strike valley were formed on north-dipping Paleozoic sedimentary rocks with dips that ranged from $<1°$ to $15°$. Much of the strike valley was constructed in the soft Permian Hermit Formation. The system was coeval with early latite generation but was clearly pre-Basin and Range faulting that was Middle to Late Miocene and Pliocene in the Verde Valley region. Similar strike valleys were noted by Lucchitta (1989) along the south margin of the Colorado Plateau west of the study area. Therefore, the strike of the southern margin of the Colorado Plateau was established by the Late Oligocene.

Two other hypotheses have been presented to explain the relations of the Beavertail Butte formation. Elston and others (1974) suggested that Verde Valley has been present since the Eocene and that Eocene gravels filled the valley and overspilled onto the Mogollon Rim. Thus Beavertail Butte and Rim gravel deposits are broadly coeval. Data above show this hypothesis to be untenable. There is no known tectonic mechanism to generate such an Eocene or Oligocene "valley" and subsequently fill it to overflowing. Such valley fill, if it did exist, would likely be mudstone and carbonate, material similar to deposits that filled part of Verde Valley in the Pliocene.

A second hypothesis (Potochnik, personal communication, 2010) suggested that the Beavertail Butte Formation is part of the Rim gravels and was subsequently faulted to its present elevation and structural position by Basin and Range graben systems. Though not totally disproven by our studies, such a series of events would require significant fault-throw reversals (~1000m) on phantom faults in which Laramide reverse faulting was exactly compensated by Miocene-Pliocene Basin and Range normal faults. Perhaps the most difficult obstacle for this hypothesis to overcome is the well-exposed onlap to the north of the Beavertail Butte formation onto Paleozoic redbeds. To permit reversal of drainage, the onlap would demand initial tilt to the north coupled with large reverse faults, up to the north, followed by present tilting to the south and subsequent down-faulting to the south. The exceptionally well-exposed Permian rocks of the Sedona area simply do not permit such tilting or structural complications. Laramide reverse faulting, which clearly exists in the Sedona area (Holm, 2001), cannot by itself explain the present stratigraphic relations of Cenozoic sedimentary rocks. Furthermore, age dates from the Beavertail Butte formation are much younger than the youngest-dated Rim gravels.

Do the Cenozoic events along and below the Mogollon Rim have a bearing on Oligocene and Miocene events in Grand Canyon? Late Oligocene to Middle Miocene sedimentary rocks are extremely rare in Grand Canyon or if they do exist, they have escaped detection. Perhaps the better-documented Oligocene-Miocene events in Central Arizona provide a proxy for Grand Canyon events. The events in Verde Valley and along the Mogollon Rim document a nibbling and headward erosion of streams into the southern Colorado Plateau, perpendicular to strike valleys; this has occurred since *ca.* 30 Ma. The two largest of these systems provide the current routes for I-17 and I-40 across the Mogollon Rim (fig. 1). The geomorphic trends in Central Arizona continue NW into the Grand Canyon region. The preserved geomorphology of Oligocene stream systems documents strong, local relief and a prominent strike valley system parallel to the Rim; most were carved in Permian redbeds, especially the Hermit Formation. Lucchitta (1989) discussed the role of strike-valley development along the southern margin of the Colorado Plateau and into the Grand Canyon region. The present Mogollon Rim continues NW of Sedona through the Aubrey Cliffs and NW to the Hualapai Plateau and across the Mio-Pliocene Hurricane fault and the Colorado River to the Shivwits Plateau. It seems probable that Oligocene strike valleys defined a paleo-Rim in the Grand Canyon region and that side canyons nibbled into the Colorado Plateau and subsequently cut headward. It also seems likely that Oligocene topography is partly reflected in the present course of the Colorado River and Grand Canyon. For instance, the NW-trending Lower Grand Canyon parallels the face of the Shivwits Plateau; it probably acquired its present course as a strike valley in the Hermit Formation *ca.* 1000 m above its present level. South of the Shivwits, Lucchitta and Jeanne (2001) calculated a cliff-retreat rate of 0.6 km/m.y. along the Kaibab-capped escarpment of the Hualapai-Shivwits Plateau region; Ranney (1988) calculated a similar rate for the retreat of the Kaibab-capped Mogollon Rim north of Beavertail Butte, using, incidentally, the same methods used by Lucchitta and Jeanne. Compared to cliff retreat rates in Mesozoic rocks, this rate is extremely slow (Lucchitta and Jeanne, 2001). Consequently, we propose that the Mogollon Rim (*senso latto*) stretched across Arizona from SE to NW and during the middle Cenozoic formed a near-stationary barrier to drainage systems attempting to escape the central Colorado Plateau. This barrier was established in the Oligocene and not fully breached until the Late Miocene, *ca*. 6 Ma. Then, a stream nibbling into the escarpment finally tapped the interior drainage of the Colorado Plateau and the Grand Canyon was carved.

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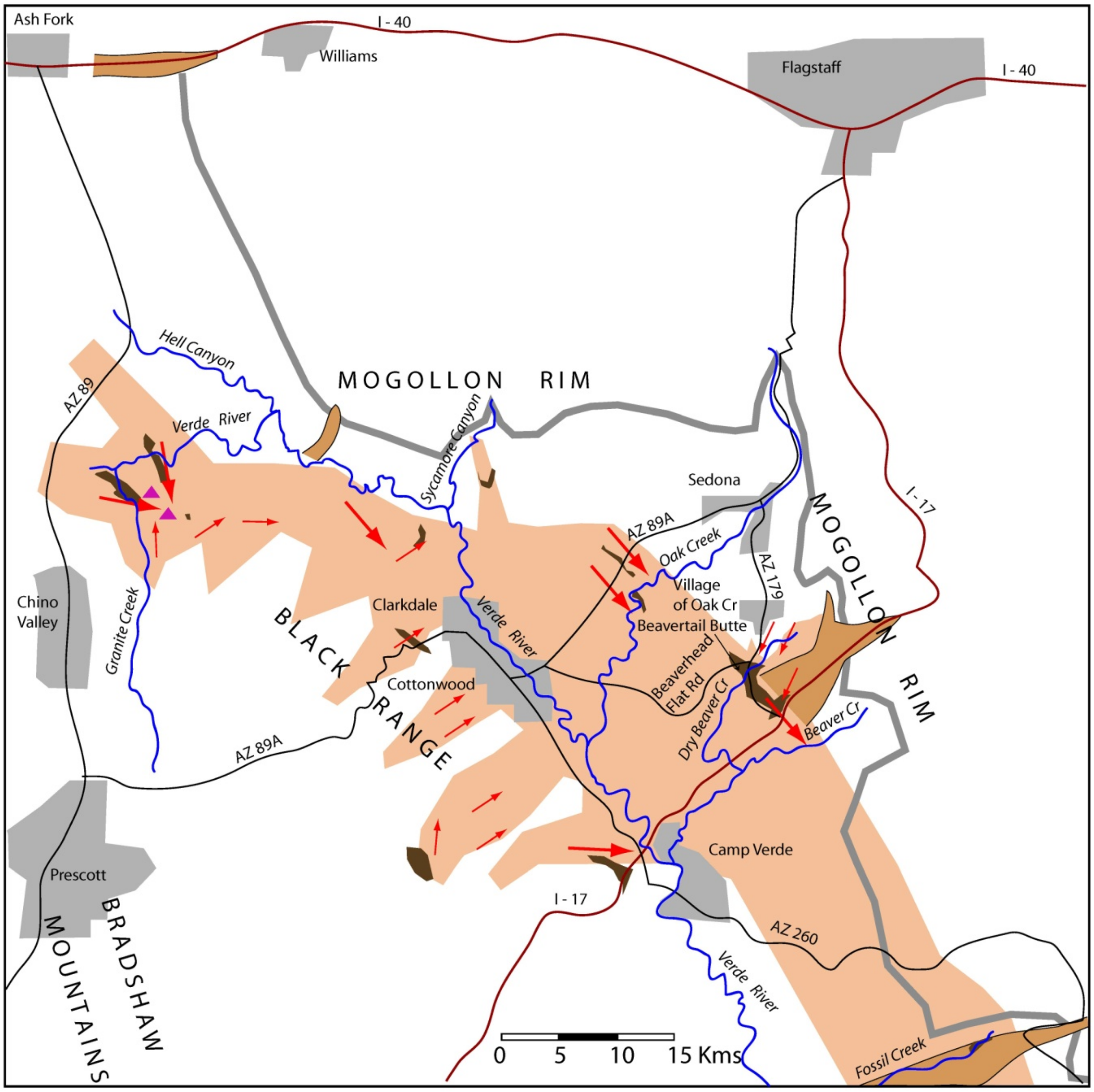

Figure 1. Index map of Central Arizona showing features discussed in text. Key to symbols and colors: dark brown – major outcrops of Little Pig (upper) member of Beavertail Butte formation and correlative units; large red arrows – paleocurrent directions for deposits believed to be major trunk streams; small red arrows – paleocurrent directions for units believed to be tributary streams; pink triangles – location of *ca.* 26 Ma Sullivan Buttes Latite; orange – Oligocene to Early Miocene paleocanyons carved perpendicular into Mogollon Rim; tan – hypothetical Oligocene braided stream system that occupied Rim-parallel strike valley with location of tributary streams also shown. The Oligocene strike valley system was mostly sourced by the paleo Black Range and Bradshaw Mountains where various Proterozoic and lower Paleozoic lithologies dominated; the system presumably drained into Tonto Basin, approximately 40 km SSE of Fossil Creek. (Modified from Loseke, 2004).

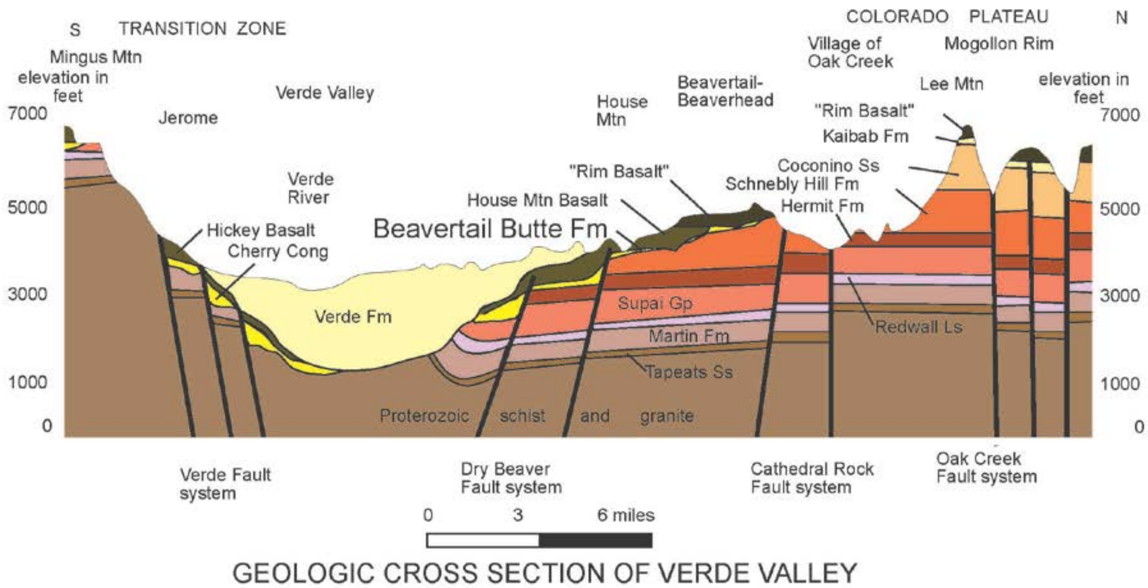

Figure 2. Simplified cross section of Verde Valley through Beavertail Butte area. Beavertail Butte formation in bright yellow. Note stratigraphic, structural, and topographic relations of the unit as discussed in text. Prior to Miocene faulting, the correlative units shown in bright yellow across the diagram formed a continuous stratigraphic unit. Hickey basalts are *ca.* 13-15 Ma; “Rim basalts” are *ca.* 6-8 Ma. (Modified from Loseke, 2004).

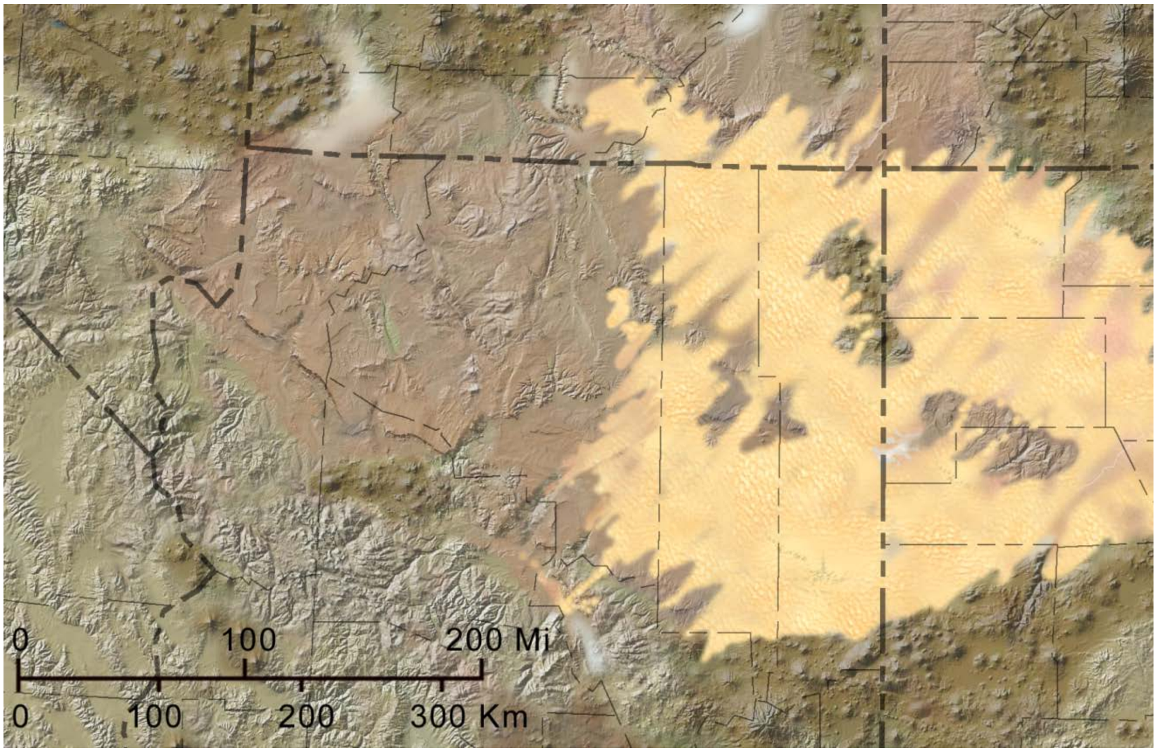

Figure 3. Hypothetical Oligocene paleogeography of Colorado Plateau and vicinity showing strike valley development along Mogollon Rim. Headward erosion into Rim carved several prominent canyons, two of which later filled with Late Miocene volcanics and provide the routes of I-40 and I-17 (see fig. 1). Note postulated continuation of Mogollon Rim into NW Arizona. Strike valley in NW Arizona may have provided opportunistic path for Pliocene-Recent Lower Grand Canyon. Tributary to NE may have carved broad valleys in Toroweap region. Extent of Chuska erg suggested by Cather, and others (2008).

Late Oligocene-early Miocene Deep Erosion on the Southern Colorado Plateau and the Southern Great Plains

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It is clear that a major episode of erosion has affected much of the southwestern USA since the late Miocene (McMillan and others, 2006). Less appreciated is evidence for earlier deep erosion on the southern Colorado Plateau and southern Great Plains during the late Oligocene–early Miocene. The recognition of an early phase of erosion on the southern Colorado Plateau was enabled by a datum provided by the paleogeographic reconstruction of the top of a large Oligocene sand sea (the Chuska erg; fig. 1), The Chuska erg was deposited ~33.5–27 Ma, and the reconstructed top of the erg would lie at a present elevation of at least 3000 m in the central Colorado Plateau (see discussion of the erg-reconstruction methodology in Cather and others, 2008).

Near the valley of the Little Colorado River and its southeastern tributaries in New Mexico, major erosion occurred after eolian deposition ended at ~27 Ma and before deposition of the Bidahochi Formation. The Bidahochi Formation (~16–6 Ma, Dallege and others, 2001, 2003) of Arizona and correlative Fence Lake Formation of New Mexico (McIntosh and Cather, 1994) are inset below the reconstructed erg top (fig. 2). Near Winslow, Arizona, pre-Bidahochi erosion removed at least 1230 m (fig. 3), using a conservative estimate of the elevation of the reconstructed erg top. Post-Bidahochi erosion was ~520 m. Near Escondida Mountain in west-central New Mexico (point A of cross-section line on fig. 1), the base of the Fence Lake Formation (middle to upper Miocene; McIntosh and Cather, 1994) is ~300 m beneath the top of the erg deposits (Chamberlin and Harris, 1994, Chamberlin and others, 1994). Post-Fence Lake erosion (post ~6 Ma) was ~150 m. Thus in both areas, late Oligocene–early Miocene erosion exceeded post-Miocene erosion by a factor of two or more. Major early Miocene erosion has also been demonstrated using apatite (U-Th)/He cooling data in the Little Colorado River valley (Flowers and others, 2008) and in the Grand Canyon area (J.P. Lee, 2010, this volume). No stratigraphic or apatite (U-Th)/He evidence exists for deep erosion the northern Colorado Plateau prior to the late Miocene.

Some aspects of the late Oligocene–Miocene geologic history of the southern Great Plains are similar to that of the southern Colorado Plateau. The Great Plains are capped with remnants of the middle to upper Miocene Ogallala Formation (or Group), a thin fluvial and eolian succession that was derived from the uplands to the west and was deposited at much the same time (~17.5–5 Ma in eastern Wyoming, ~12–5 Ma in eastern New Mexico; McMillan and others, 2002; Gustavson, 1996) as the Bidahochi Formation of the Colorado Plateau. The age of the subcrop beneath the Ogallala Formation is illustrated in figure 4, a north-south cross section on the Great Plains along 103.5° W longitude. Near the “gangplank” (or Cheyenne tablelands) east of the Laramie Range, the Ogallala Formation overlies the lower Miocene beds of the Arikaree Formation with no more than a few million years of depositional hiatus. The lacuna represented by the basal Ogallala unconformity increases southward as it oversteps progressively older Paleogene and Mesozoic strata. In southeastern New Mexico, the Ogallala Formation unconformably overlies Triassic and, locally, Permian beds. There, regional stratigraphic relationships indicate uplift followed by 1.0–1.5 km of erosion occurred prior to deposition of the Ogallala Formation, and the sub-Ogallala unconformity represents a lacuna of ~230–250 m.y.

The southward beveling of older strata beneath the Ogallala on the southern Great Plains is largely the result of erosion since the late Oligocene. This is most clearly illustrated by the geologic relationships near the Capitan intrusion (fig.1), a 28.3 Ma large granitic stock/laccolith in southeastern New Mexico (Allen and McLemore, 1991). The Capitan intrusion attains an elevation of ~3100 m, and an estimated 1.2–1.6 km of strata have been eroded from above it (Phillips, 1990). About 20 km south of the Capitan intrusion near Fort Stanton, pediment gravels correlated to the Ogallala Formation by Kelley (1971)(unit QTg of Rawling, 2008) crop out at ~2000–2200 m elevation. Thus, following intrusion of the Capitan pluton at 28.3 Ma, at least 2 km of erosion occurred before deposition of the Ogallala Formation began in the middle to late Miocene. By contrast, maximum post-Miocene (post-Ogallala) incision in the Pecos River valley to the east is ~500 m. Late Oligocene–early Miocene (pre-Ogallala) erosion in southeastern New Mexico thus exceeds late Miocene–Recent (post-Ogallala) erosion by a factor of 2–4. The magnitude of pre-Ogallala uplift and erosion diminishes northward on the Great Plains; at the latitude of northern Colorado and Wyoming the post-Ogallala erosional event is predominant (e.g., McMillan and others, 2002; 2006).

Voluminous sediment eroded from the Great Plains during the late Oligocene–early Miocene was probably deposited in the Gulf of Mexico. The fate of sediment shed from the southern Colorado Plateau during this time interval is less certain. It is possible that this sediment was stored on the northern Colorado Plateau, but no evidence for a depositional episode of appropriate volume is preserved. Rivers probably did not exit the Colorado Plateau to the south or west until extensional collapse of former Laramide highlands in these directions allowed for drainage reversal. In the case of the paleo-Salt River in Arizona, this reversal did not occur until the middle Miocene (Potochnik and Faulds, 1998, Potochnik, 2001). It is possible that rivers draining the southern Colorado Plateau during the late Oligocene–early Miocene exited the plateau eastward to the Gulf of Mexico, before the ~16 Ma beginning of rapid subsidence and rift-shoulder uplift in the Rio Grande rift (Chapin and Cather, 1994) became a major impediment to transverse rivers.

It is perhaps more likely that late Oligocene–early Miocene rivers exited the Colorado Plateau to the north. Pre-Ogallala southward beveling of the Great Plains from east-central Colorado to southeastern New Mexico implies a slight northward tilting of this broad region. It is possible that the Colorado Plateau experienced a similar slight northward tilting, resulting in erosion of the southern part of the plateau and development of north-flowing drainages. Although the pre-Bidahochi paleovalleys on the southern Colorado Plateau clearly sloped to the north (Love, 1989; W.R. Dickinson, this volume, 2010), no evidence of north-flowing rivers of late Oligocene–early Miocene age on the northern Colorado Plateau is known. However, the Bidahochi Formation that began to aggrade in paleovalleys of the southern Colorado Plateau during the middle Miocene contains fossil fish with affinities to fish in rivers of the Pacific Northwest, implying a hydrologic connection to that region (Spencer and others, 2008).

The cause of late Oligocene–early Miocene uplift and deep erosion on the southern Colorado Plateau and the southern Great Plains is unclear. The broad extent of the uplift, as shown by the ~1000-km, north-south extent of significant pre-Ogallala erosion on the Great Plains, is suggestive of a buoyancy source in the upper mantle. A potential source uplift is mantle buoyancy resulting from basalt extraction by partial melting of the mantle during Oligocene ignimbrite flare-up (IFU) volcanism, as has been proposed for the eastern Colorado Plateau by Roy and others (2004). The region of the most voluminous IFU volcanism, however, was in the Sierra Madre Occidental, which lies south of the area of late Oligocene–early Miocene uplift and deep erosion on the southern Colorado Plateau and the southern Great Plains (fig. 5). In the Sierra Madre Occidental, ~300,000 to 500,000 km^3 of silicic ignimbrite was erupted ~46–15 Ma (Cather and others, 2009). Such volcanism requires basalt extraction from large volumes of mantle (>45 million km^3; Farmer and others, 2008), and thus may modify the buoyancy of the mantle over a large area. Note also that the Ogallala Formation and underlying strata are broadly arched over the Jemez volcanic lineament (fig. 4). Uplift related to Miocene and younger magmatism along the Jemez lineament has been interpreted by Wisniewski and Pazzaglia, (2002) to have influenced the late Cenozoic incision history of the Canadian River (about 70 km west of the line of section of fig. 4.

The Oligocene–Miocene is the least understood aspect of the Cenozoic drainage evolution of the Colorado Plateau. During the Laramide orogeny, rivers entered the Colorado Plateau from the south and west. Eocene rivers on the southeastern Colorado Plateau exited the plateau to the east and flowed across the Great Plains region toward the Gulf of Mexico. Paleogene rivers on the remainder of the Colorado Plateau terminated in closed basins in the northern part of the plateau. A major river that entered the southwestern Colorado Plateau near the present Grand Canyon area headed on the continental divide in eastern California (the California River of Wernicke, 2009) and terminated in the southern Uinta Basin (W.R. Dickinson, 2010, this volume). A profound episode of drainage reorganization occurred during the Oligocene–middle Miocene, during a time of major volcanism and erg development on the Colorado Plateau and extensional collapse of Laramide highlands south and west of the plateau. By latest Miocene time, following the integration of Colorado Plateau drainages to the lower Colorado River by capture/spillover at the east Kaibab monocline, the continental divide had been reestablished far to the east, on the eastern part of the Colorado Plateau.

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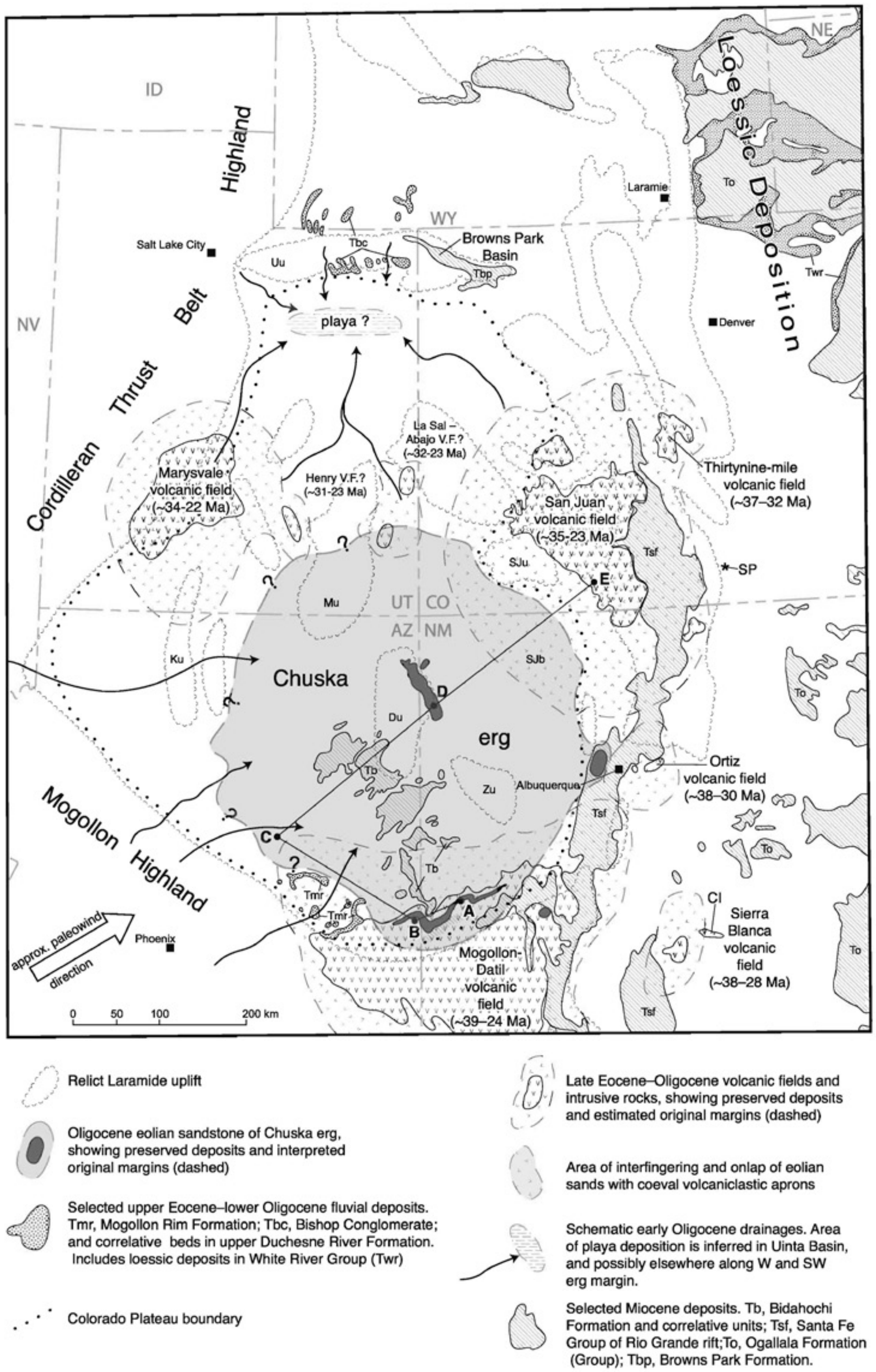

Figure 1. Map showing interpreted early Oligocene paleogeography in the Colorado Plateau–Rocky Mountain area in relation to relict Laramide uplifts and selected Miocene sedimentary deposits. A–E is line of section for figure 2. Selected Laramide uplifts are: Du, Defiance uplift; Zu, Zuni uplift; Ku, Kaibab uplift; Mu, Monument uplift; SJu, San Juan uplift; Uu, Uinta uplift. Selected intrusions are CI, Capitan intrusion; SP, Spanish Peaks. Modified from Cather and others (2008).

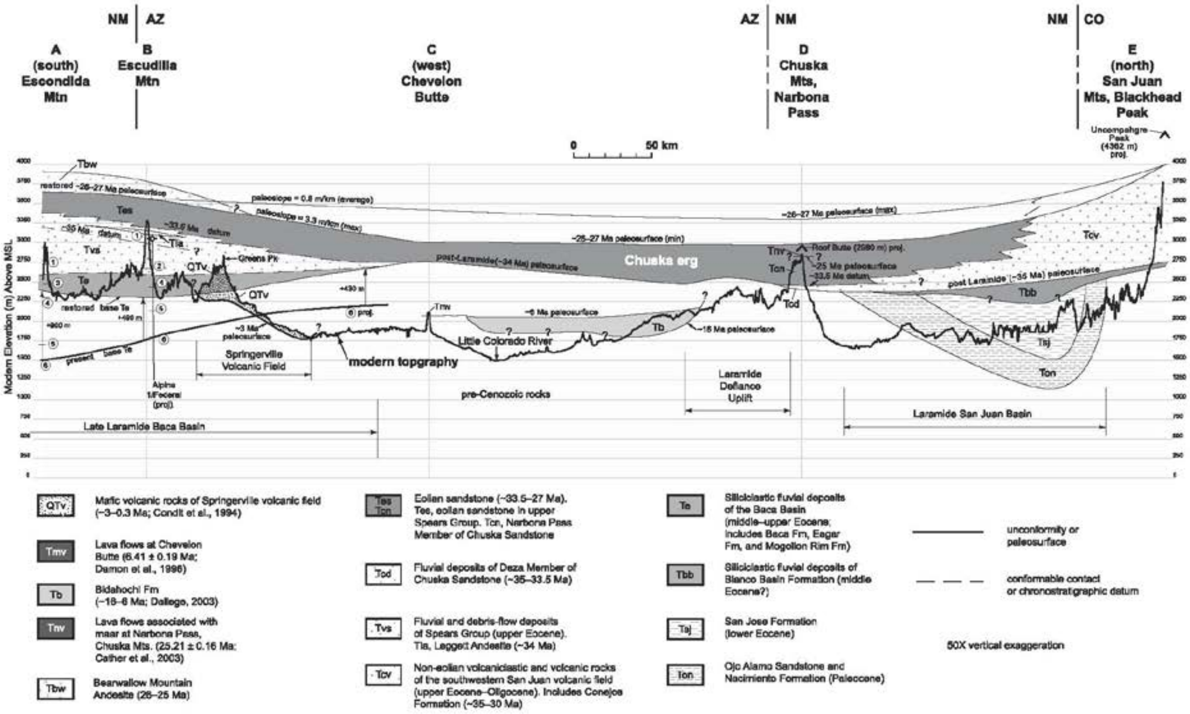

Figure 2. Regional cross-section showing present-day elevations of reconstructed stratigraphic successions of Paleocene to Pleistocene age in relation to modern Colorado Plateau topography. Line of section is shown in figure 1. Note that the elevation of strata in the southern (left) part of the cross-section has been restored (vertical arrows above present base of Te) to account for structural lowering of the southern Colorado Plateau margin [see Cather and others (2008) for details of erg reconstruction]. Unrestored elevations of top of eolian succession at (1) Escondida Mountain and (2) Escudilla Mountain shown by tic marks and circled numerals. Approximate ~35 Ma chronostratigraphic datum, given by base of volcaniclastic unit of Cañon del Leon (35.6 Ma; Chamberlin and Harris, 1994) at Escondida Mountain and by the Bishop Peak Tuff (35.1 Ma, projected from nearby Alpine1/Federal well) at Escudilla Mountain; A range of elevations for the reconstructed ~27–26 Ma top surface of the Chuska erg is depicted, using end-members based on steep vs. average slopes of modern ergs. The topographically highest points in the Chuska Mountains and San Juan Mountains are projected from areas north of the line of section. From Cather and others (2008).

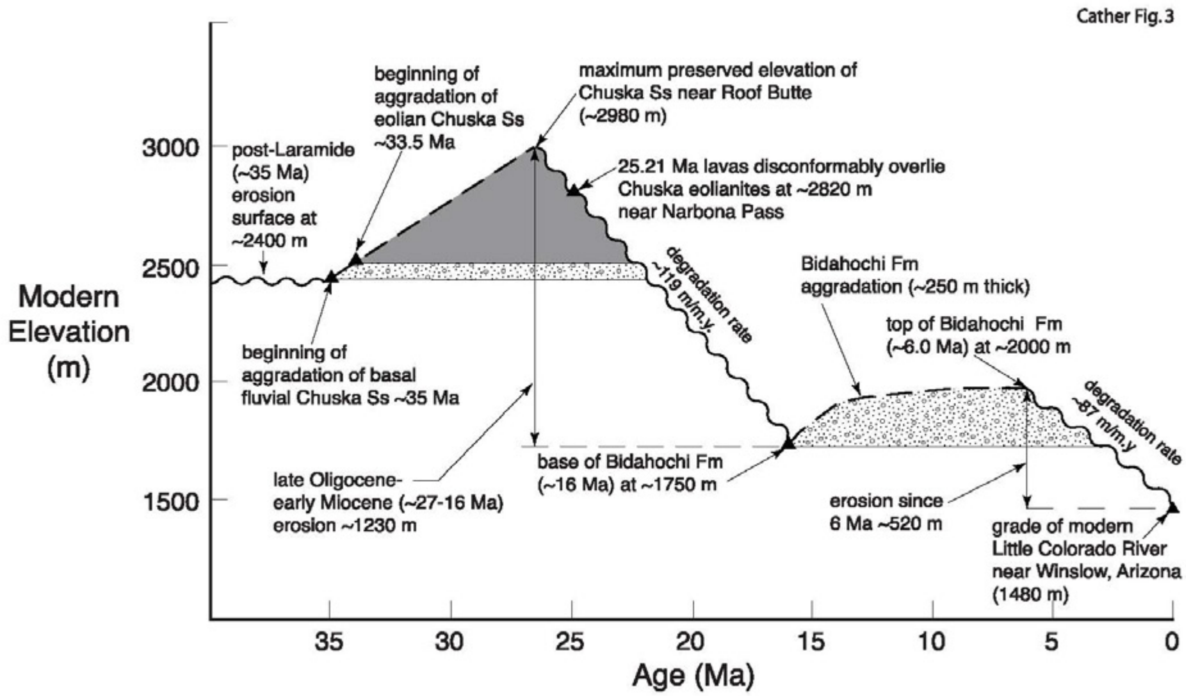

Figure 3. Late Eocene to Recent aggradation and exhumation history of the southern Colorado Plateau relative to present-day elevations, based on a conservative (minimum) estimate of elevation of Chuska erg surface. Note that the Chuska Sandstone and Bidahochi Formation outcrops are ~100 km apart. From Cather and others (2008).

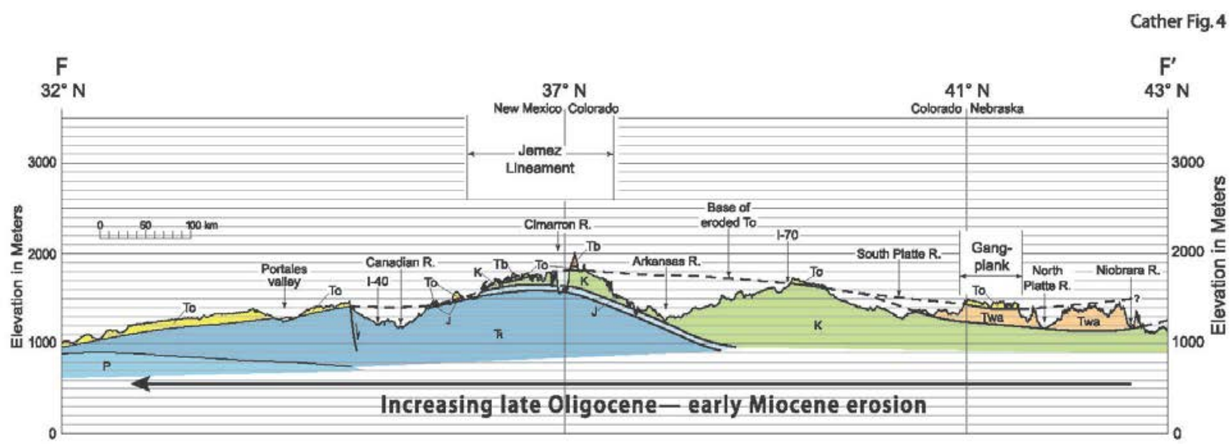

Figure 4. North-south cross section from 32°N to 43°N along 103° 30', parallel to, and approximately 150 km east of, the Rocky Mountain front. Vertical exaggeration is 100x. Stratigraphic units are: P, Permian; Tr, Triassic; J, Jurassic; K, Cretaceous; Twa, White River Group (upper Eocene-Oligocene) and Arikaree Formation (lower Miocene); To, Ogallala Formation (middle to upper Miocene); Tb, basalt flows (upper Miocene). Note that the Ogallala Formation (or Group) oversteps progressively older strata to the south. Also note the Ogallala Formation is broadly arched across the trend of the Jemez volcanic lineament, as are subjacent Mesozoic strata. Line of section is shown on figure 5.

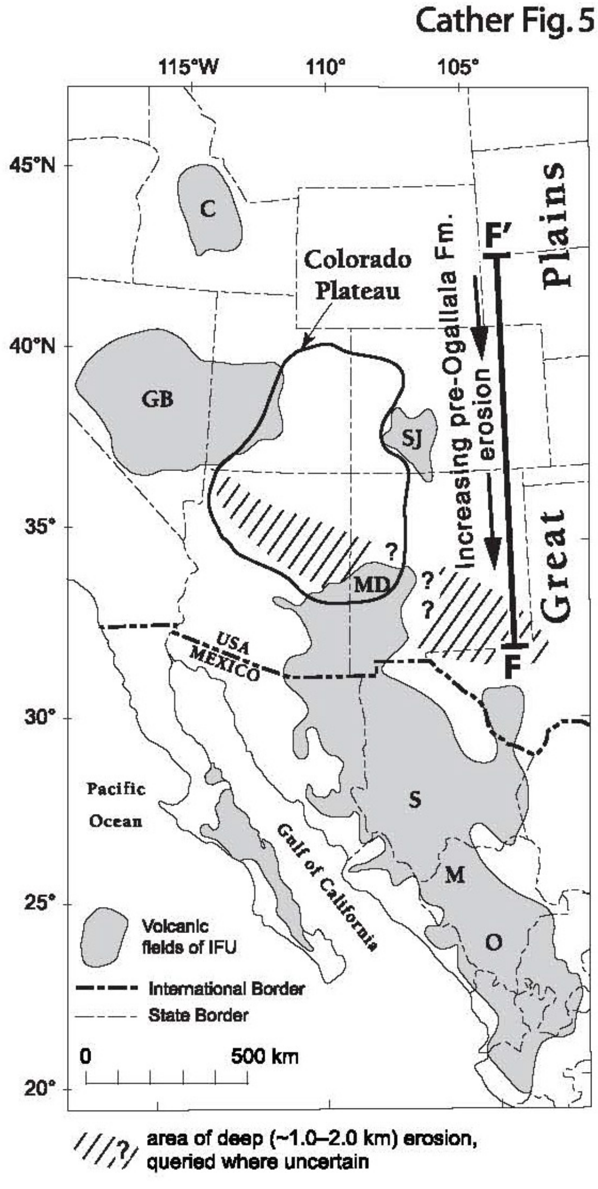

Figure 5. Map of areas of late Oligocene-early Miocene deep erosion on the Colorado Plateau and Great Plains relative to volcanic fields of the ignimbrite flare-up (IFU) in Mexico and southwestern USA. SMO, Sierra Madre Occidental; MD, Mogollon–Datil, New Mexico; GB, Great Basin; C, Challis, Idaho; SJ, San Juan and central Colorado volcanic fields. F–F' is line of section for figure 4. Modified from Cather and others, 2009).

Significance of the Grand Mesa Basalt Field in Western Colorado for Defining the Early History of the Upper Colorado River

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Grand Mesa exists today because it is capped by a basalt sequence (figs. 1 and 2) that ranges in age ($^{40}Ar/^{39}Ar$) from 9.45 to 10.99 Ma (mean = 10.17 Ma; fig. 3). The basalt sequence has a present-day surface area of about 155 km^2 (60 mi^2); however, the original field may have been as much as 907 km^2 (350 mi^2), or greater. The largest surviving part of the field has a "Y-shaped" outline, with Crag Crest forming the stem, and the Palisade and Flowing Park lobes forming the branches. Additional outliers of the field exist to the east, including Leon Peak, Green Mountain, Priest Mountain, Mt. Hatten, Crater Peak, and Mt. Darline. A broad expanse of detached basalt forms a landslide bench around the basalt remnants. This mass wasting was produced mainly by Pleistocene ice loading (Yeend, 1969; Baum and Odum, 1996). The basalt rests on a paleo-topographic Miocene (?) surface that currently ranges in elevation from 3,422 m (11,277 ft) on the east (Crater Peak) to 2,936 m (9,632 ft) on the west (Shirt Tail Point). This surface has an average east-to-west gradient of about 9.5 m/km (50 ft/mi) over a lateral distance of 51 km (32 mi) (fig. 2); however, the gradient is not uniform. Between Crater Peak and Leon Peak, the gradient is about 8 m/km (43 ft/mi), whereas between Crag Crest and the join of the Palisade and Flowing Park Lobes it is approximately 18 m/km (96 ft/mi), and on the west end of the Palisade Lobe it is about 4 m/km (19 ft/mi). It is possible that these gradient variations document post-eruption differential uplift (or subsidence). Alternatively, the steep-gradient area just west of Crag Crest may represent bulging produced by intrusion of basaltic magma. It should be noted that post-eruption tilting of the basalt cap was first suggested by Young and Young (1968).

Up to 26 basalt flows have been observed in western Grand Mesa, where the total thickness ranges from 64 to 187 m (209 to 613 ft), with thinning from east to west. Individual flows in the western area range in thickness from 1.1 to 21.6 m (3.7 to 70.9 ft). For eastern Grand Mesa, the number of flows present at a given locality ranges from three to eight. In contrast, the total thickness of the eastern sequence is only 15 to 34 m (50 to 110 ft). Regardless of location, individual flow remnants are lenticular and rarely more than 90 m (295 ft) wide. Interflow sediment beds are locally present in western Grand Mesa, but cannot be precisely correlated. Examination of stretched vesicles (N = 2,714) shows that the average flow movement was to the west and southwest (vector mean = 267º), which is consistent with the slope of the sub-basalt surface.

Currently, twenty-three $^{40}Ar/^{39}Ar$ age dates exist for the Grand Mesa area, as summarized in figure 3. Data for the Lands End and Skyway areas of the Palisade Lobe are from Kunk and others (2002), and the single date from the Bowie Creek sill (coal mine near Sommerset, CO) is from Robeck (2005). The remaining dates are from New Mexico Technological University (funding from the Colorado Rockies Seismic Experiment and Transects, CREST, Project). The values range from 9.45 to 10.99 Ma (1.54 Ma), with a mean of 10.17 Ma. When grouped geographically, these data suggest that the eastern Grand Mesa samples (N = 7) are slightly older (mean = 10.25 Ma) and have a narrower distribution (range = 0.31 Ma) than those from western Grand Mesa (N = 16, mean = 10.14 Ma, range = 1.54 Ma), which have a broader distribution.

Field evidence suggests that emplacement of the basalt probably occurred in two stages and from two separate vent areas. The initial eruptions presumably came from the "Lombard" vent (currently represented by two east-west oriented basalt dikes; see fig. 1) at the southeastern edge of the basalt field. Flows from this vent probably followed a series of paleo-valleys on the northern flank of the Oligocene West Elk volcano. The second eruption sequence occurred at the "Lily Lake" vent, which is documented by a north-south oriented basalt dike and scattered remnants of pyroclastics. The Lily Lake dike clearly crosscuts older flows, which may have erupted from the Lombard vent.

Underlying the basalt sequence is a Miocene (?) sequence of poorly consolidated mudrock, lithic sandstone, and conglomerate, called the "unnamed unit" by Yeend (1969), and herein called the "Goodenough" formation (proposed type section at Goodenough Reservoir at the base of Crater Peak). The age of the Goodenough formation is not constrained, but may be equivalent to Browns Park, North Park, or Troublesome Formations (Oligocene-Miocene; Izett, 1975). The sedimentologic and stratigraphic characteristics of the Goodenough formation are poorly understood because it is obscured by heavy vegetation and has been significantly disrupted by mass wasting (Baum and Odum, 1996). Outcrops exist throughout the landslide bench (fig. 1), plus it was partly penetrated (several meters at best) in a series of core holes drilled by the U.S. Bureau of Reclamation on the Palisade lobe (Weston, 1987). It should be noted that at the extreme western edge of the Palisade lobe, the basalt sequence rests on polylithic gravel from the ancestral Colorado River (Czapla and Aslan, 2009; see fig. 2).

The Goodenough formation rests unconformably on the Paleocene-Eocene Green River and Uinta Formations, and is up to 274 m (900 ft) thick (Yeend, 1969). In central and eastern Grand Mesa, the Goodenough can be subdivided into two intervals. The lower interval (two-thirds of the total thickness) consists of multi-colored, lacustrine (?) mudrock and limestone, whereas the upper part consists of fluvial sandstone, conglomerate, and mudrock. In many locations the fluvial facies is immediately overlain by basalt. Clasts in the fluvial facies are dominated by diorite and andesite from the West Elk Mountains to the southeast and probably the Elk Mountains to the east. Rare (exotic) clasts of quartz, K-feldspar, quartzite, marble, granite, gneiss, and schist also occur, suggesting additional sediment contributions from higher uplifts to the east (White River, Elk, Sawatch, and Gore ranges). It is likely that some of the exotic clasts are reworked from the Permian Maroon Formation.

The Gunnison River and North Fork of the Gunnison River have incised between 1,488 and 1,775 m (4,881-5,824 ft) since emplacement of the basalts. It is possible to calculate the incisions rates of the former rivers from the aerial distribution of recent $^{40}Ar/^{39}Ar$ dates and the current elevations of the sub-basalt strath surfaces. On the western edge of the field (data from Skyway, Flowing Park, and Lands End), the incision rates range from 135 to 168 m/my (0.44 to 0.55 ft/ky), with the total incision (i.e., relief) ranging from 1,488 to 1,771 m (4,881-5,810 ft). In the middle of field (Leon Peak), the incision rate is 169 m/my (0.55 ft/ky), with a total incision of 1,751 m (5,746 ft), whereas on the eastern edge of the field (e.g., Crater Peak) the rate is 171 m/my(0.56 ft/ky), with a total incision of 1,775 m (5,824 ft). Thus, these calculations indicate that the river-incision rates and the magnitudes of the incision are increasing from west to east.

A series of gravel-capped surfaces (now terraces) formed on Grand Mesa's southern and southwestern flanks during its late Cenozoic erosional development (fig. 4). They are carved into the Mancos, Iles, Williams Fork, Wasatch, and Green River Formations (Cole and Sexton, 1981). Geographically and genetically, these surfaces can be grouped into two types. Those north and northwest of Delta, CO, (the western terraces on figure 4) are interpreted as pediment surfaces (Baker and others, 2002; Rider and others, 2006; Cole, 2007; Darling and others, 2007), whereas those between Delta and Hotchkiss, CO, (eastern terraces) are interpreted as alluvial surfaces (Cole and Sexton, 1981; Cole, 2007) associated with large-scale sediment transport down Surface and Leroux Creeks. The alluvial surfaces have fan-shaped geometries and are primarily the result of glacial outwash (Yeend, 1969). These terraces can be subdivided into four levels. The ages of these surfaces have not been determined, except for the level-3 gravels, which occasionally include interbeds of Lava Creek B ash (640 ka) (Aslan and Cole, 2002). Based on the river-incision rates discussed above, the maximum age of the highest gravel-capped surfaces (level 2) may be as old as 4 Ma, whereas the youngest gravel-capped surfaces (level 5) may be as young and 70 Ka. The gravel sequences range in thickness from 3 to 58 m (10 to 190 ft), with clasts strongly dominated by basalt (95 to 99 percent). Sandstone, quartzite, chert, diorite, andesite, schist, gneiss, granite, and pegmatite make up the minor clast assemblage. The non-basalt clasts probably represent reworked high-elevation gravels from the ancestral Gunnison River or North Fork Rivers, or from the Goodenough formation. It is also possible that the non-basalt clasts on the western flanks of Grand Mesa are from the ancestral Colorado River (Czapla and Aslan, 2009).

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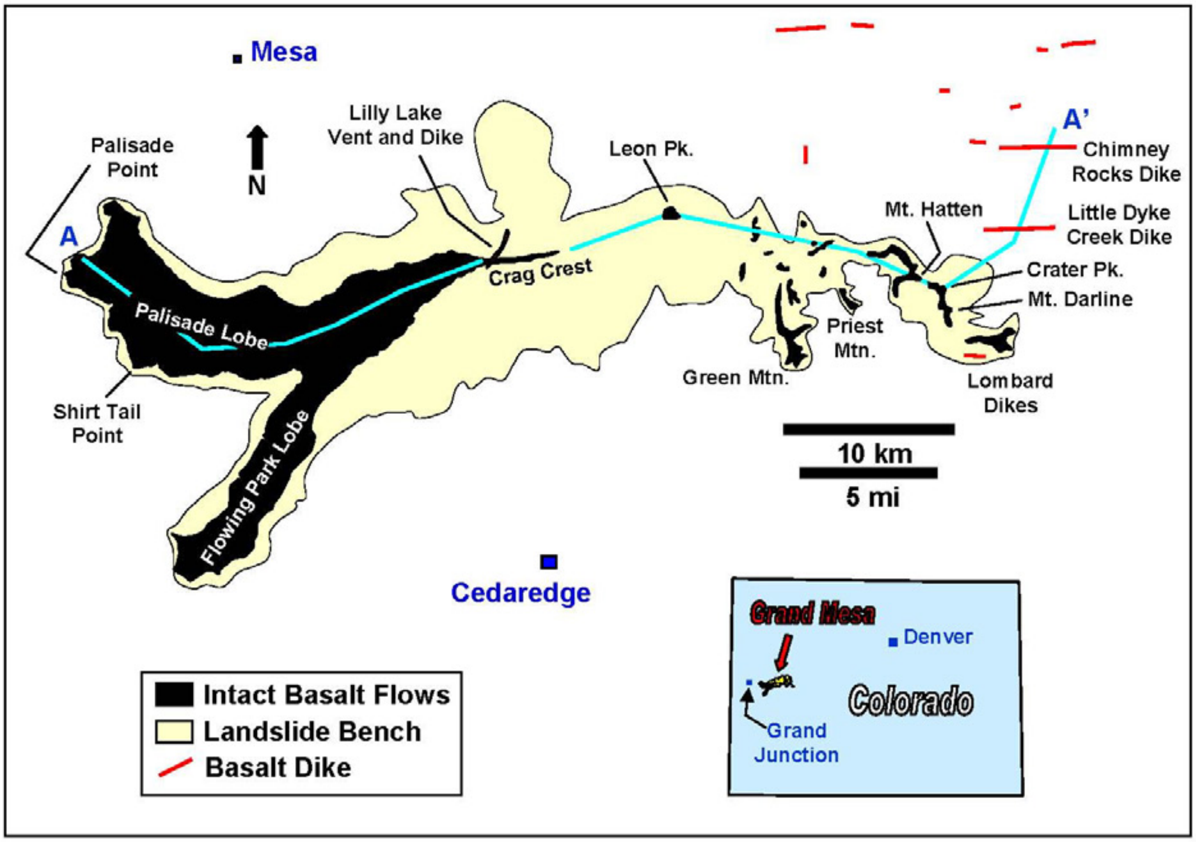

Figure 1. Index map of the Grand Mesa area, western Colorado. See figure 2 for cross section.

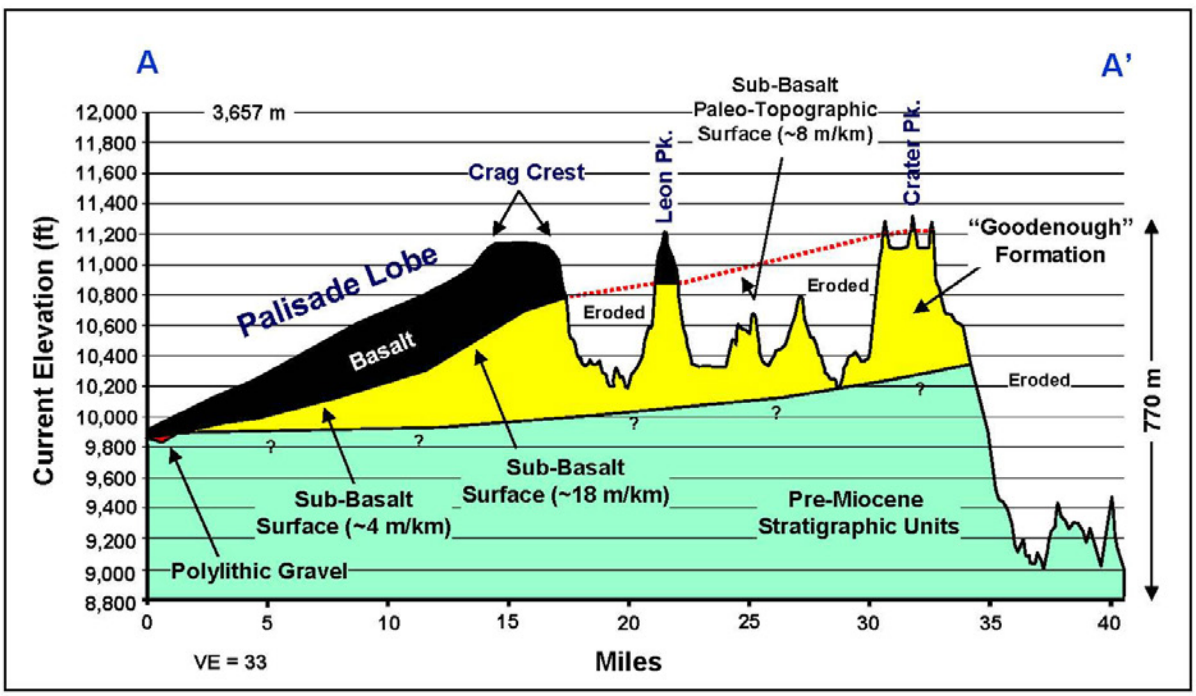

Figure 2. East-west cross section of Grand Mesa. Section line shown in figure 1.

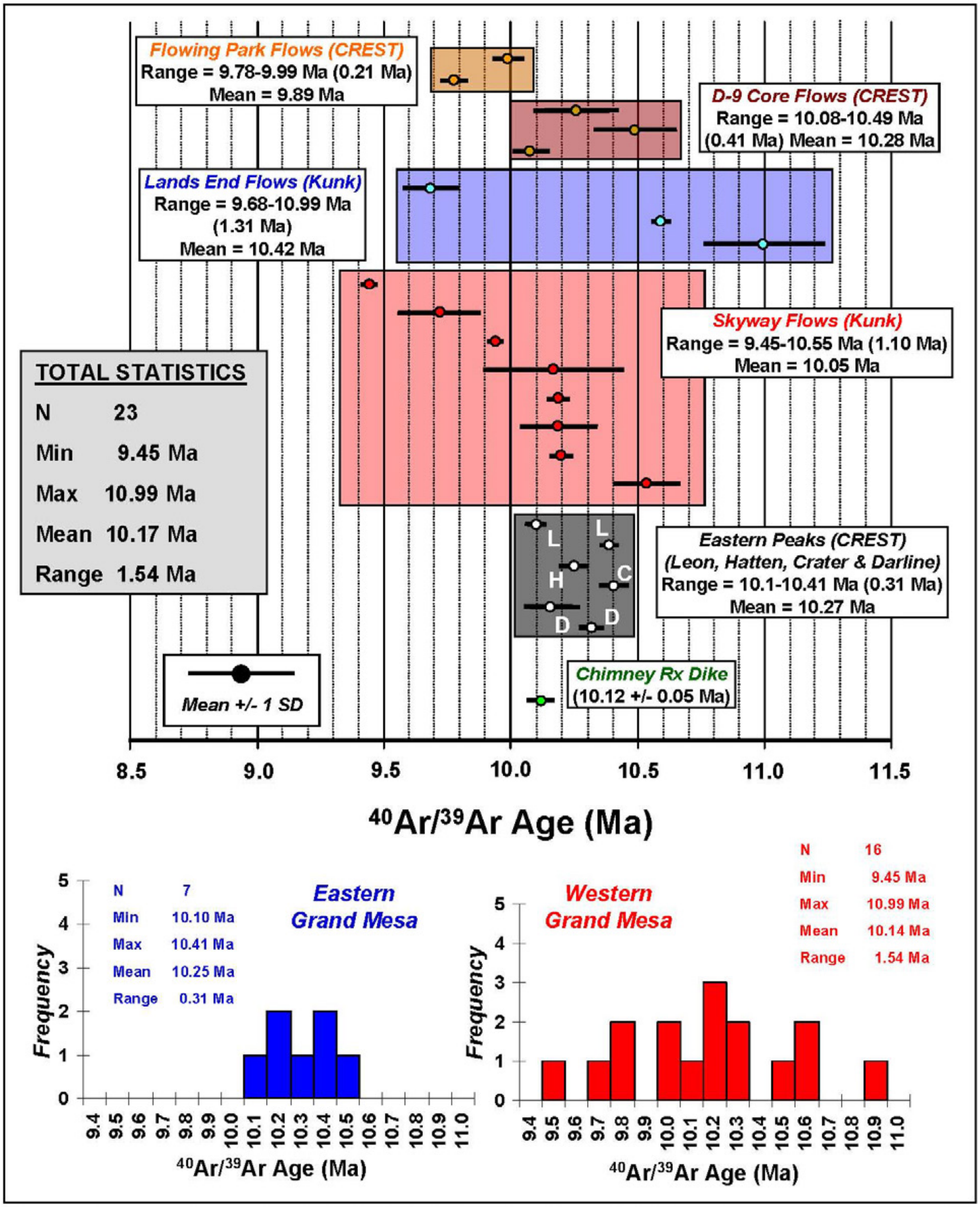

Figure 3. Summary of $^{40}Ar/^{39}Ar$ data from Grand Mesa; see figure 1 for locations. Data are from Kunk and others (2002), Robeck (2005), and unpublished analyses from New Mexico Technological University (CREST project). Data have been normalized to the FC-sanidine standard (28.02 Ma).

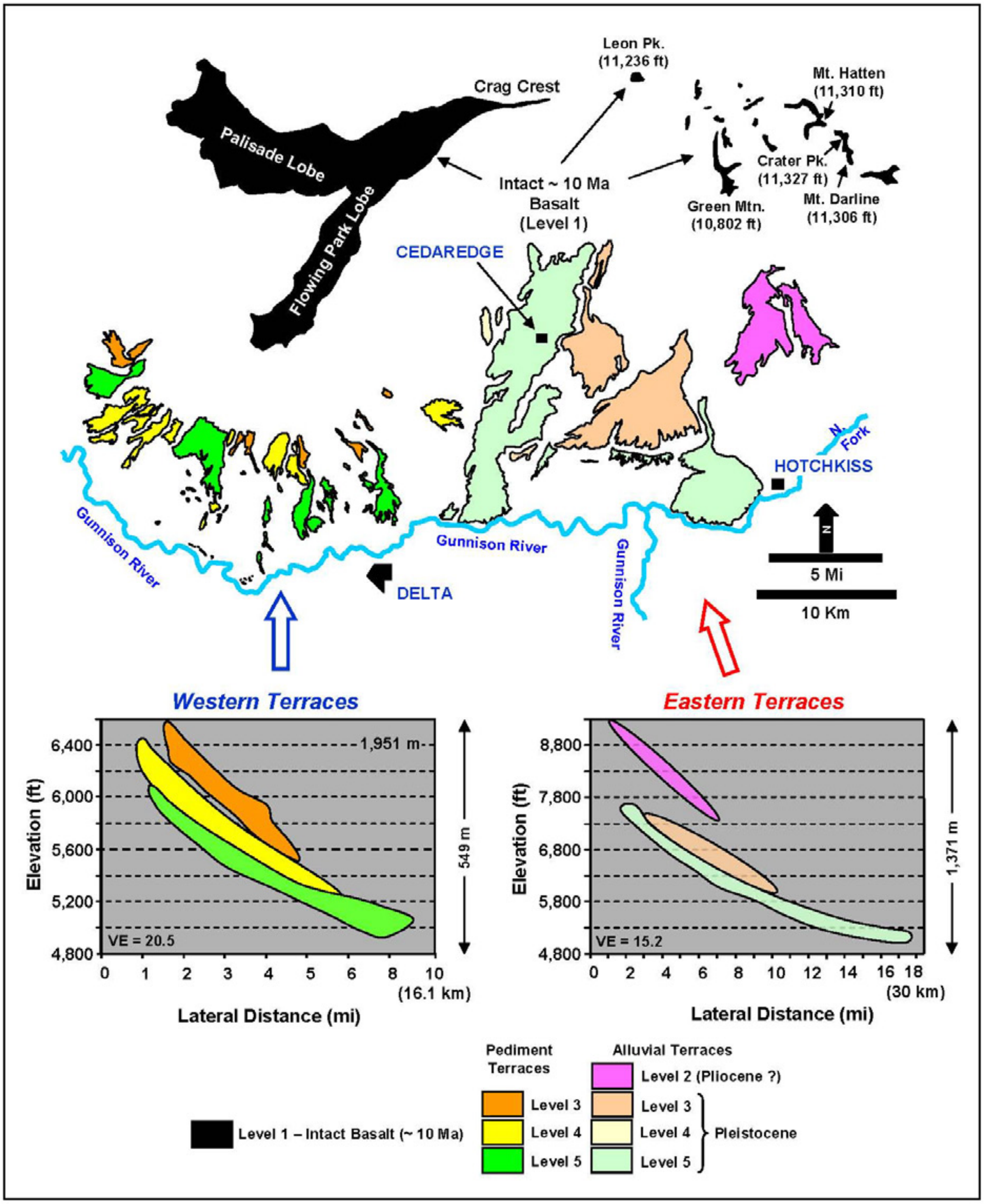

Figure 4. Map of late Cenozoic gravel-capped terraces on the south flank of Grand Mesa (upper), and summary of topographic profiles for these surfaces. Modified from Aslan and others, (2008).

Geochemistry of Springs, Travertines and Lacustrine Carbonates of the Grand Canyon Region Over the Past 12 Million Years—The Importance of Groundwater on the Evolution of the Colorado River System

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The modern Colorado River is part of a hydrologic and hydrogeochemical system that includes: 1) the river carrying far-traveled snow melt from the Rocky Mountains, 2) indigenous spring waters that form the base flow for most of the tributaries of the Colorado Plateau (e.g Havasu Creek, Little Colorado River, Salt River), and 3) a groundwater aquifer system that feeds this baseflow, including the Muav-Redwall (RM) karst aquifer system. We examine the geochemistry of modern springs and their associated travertine deposits and compare them to older carbonate deposits such as the Hualapai Limestone and Bouse Formation. The goal is to provide insight into past hydrologic and hydrochemical settings in order to better understand how the hydrologic system has evolved into the present Colorado River system.

Modern springs, groundwater, and travertines in the Colorado Plateau hydrologic system represent variable mixing of deeply derived (endogenic or "lower world") fluids with meteoric (epigenic or "upper world") water. Meteoric recharge is dominated by high elevation recharge from the San Francisco volcanic field, but also includes other high elevation regions such as the Kaibab uplift. Mixing relationships reflect balance between the much larger volumes of the "upper world" waters and the small volume but much more geochemically potent "lower world" fluids that are CO2-charged fluids, containing mantle 3He, and ascending along fault conduits (Crossey and others, 2009). We apply two main tracers to understand water/carbonate systems through time. 1) C and O stable isotopic values of water and carbonate, and 2) 87Sr/86Sr isotopic values of water and carbonate. The two systems, considered together, offer greater insight than either on its own.

Observations based on modern systems are: 1) the δ 18O value in water is strongly influenced by recharge elevation of waters; 2) the δ 13C values of dissolved inorganic carbon reflects mixing of carbon reservoirs, including: dissolved mineral carbonate from the groundwater aquifer (near 0 to +2 permil for the Redwall-Muav aquifer), biologic processes (preferentially incorporating 12C in organic matter, leading to low δ 13C values at ~ -28 permil), and endogenic contributions (near -5 permil). Carbonates that are deposited from CO2 - supersaturated waters in travertines and lakes provide a paleo hydrochemical record of water composition once fractionation factors during crystallization are factored in.

The observed slight downstream increase in 87Sr/86Sr in modern Colorado River water through Grand Canyon (fig. 1; Spencer and Patchett; 1997; Patchett and Spencer, 2001) has been explained by inputs of deeply sourced (endogenic) fluids in springs of the Grand Canyon incised aquifer system (Crossey and others, 2006). 87Sr/86Sr ratios in carbonates are a direct reflection of paleo water chemistry because Ca and Sr readily substitute in carbonate rocks. This is demonstrated by plots showing similar 87Sr/86Sr of modern waters and their coexisting travertine deposits (Crossey and others, 2006). Our conclusion based on 87Sr/86Sr from modern waters is that 87Sr/86Sr values of endogenic springs which have travelled through granitic basement of western Grand Canyon are as high as 0.0735. These spring waters mix with regional groundwaters of ~0.705-0.710 to produce the observed wide range of 87Sr/86Sr measured in modern springs and travertines of Grand Canyon (fig. 1).

The 12-6 Ma Hualapai Limestone of Grand Wash trough was studied for its chemostratigraphy based on several measured sections (see Pearce and others, this volume). The several hundred meter thick sections of lacustrine carbonate (Grand Wash Trough section reaches a thickness of 300 m) indicate long-term throughput of groundwater (non-evaporite conditions) in this lake/marsh system. Figure 2 is a C-O plot that shows that travertines have distinctive range of δ 13C values depending on local spring composition. The large variations shown in C (over 12 permil) suggest that different groundwaters record distinctive local mixing proportions (e.g. eastern versus western Grand Canyon) plus local effects of biological activity (e.g. in larger volume systems such as Havasu and Hualapai). Hualapai Limestones overlap most strongly with modern travertines being deposited in central Grand Canyon by Havasu Creek (O'Brien and others, 2006), with characteristically high δ 13C values (> 2 permil), and δ18O values of -13 to -8 permil. These stable isotope values differ from values of carbonate that would be in equilibrium with the modern Colorado River (δ 13C about -4 permil). We conclude that Hualapai Limestone was deposited by Colorado Plateau groundwaters, similar to the mix of RM aquifer waters and endogenic fluids seen in today's groundwater at Havasu Creek, but quite distinct from Colorado River water. The marked difference in stable isotope composition between Colorado River water and Hualapai Limestone argues against models for infiltration, piping, and flow of paleo Colorado River water through caves and sink holes (during Hualapai deposition) as a major step in integration of the Colorado River across the Kaibab uplift and through Grand Canyon (Hunt, 1956, Hill and others, 2007, Pederson, 2008). However, our data do not rule out arrival of Colorado River water by such mechanisms after Hualapai deposition.

The δ 18O and δ 13C isotopic values in the Hualapai Limestone, especially the Grand Wash Trough section, both show progressive up-section decrease within measured sections. This indicates that groundwater feeding the Hualapai lakes and marshes had progressively larger meteoric input through time (see Pearce and others, this volume). The time duration of this change is 12-6 Ma indicating gradual change in western Colorado Plateau groundwater chemistry, perhaps due to building of a high elevation recharge area of the San Francisco volcanic field 10-4 Ma. This gradual change is not explainable by lake spill-over models for an abrupt freshening of Lake Hualapai. Gradual decrease in 87Sr/86Sr up-section is also compatible with increased meteoric recharge in Plateau groundwaters from 12-6 Ma. Our conclusion is that the Hualapai Limestone was a spring-fed lacustrine system that records evolving regional groundwater flowpaths from the western Colorado Plateau, beginning with 17 Ma lowering of the Basin and Range on Grand Wash fault, and probably influenced by increased high elevation recharge due to building of the San Francisco volcanic field 10-6 Ma.

The Bouse Formation of the Lower Colorado River region is geochemically distinct from Hualapai Limestone (fig. 2) and reflects its own local hydrologic setting 5.5 Ma. The base of the Bouse consists of thin carbonate deposits at variable elevations in a series of lake basins, the

Mojave basin to the north and Blyth and Havasu basins to the south (nomenclature of Spencer and others, 2008). Stable isotope plots show that many points fall on an apparent mixing trend between sea water values (a possible marine endmember) and an endogenic spring endmember similar to western Grand Canyon springs (fig. 2). Thus, the simplest explanation for the combined O and C values is that this unit represents a mixture of marine, river, and spring inputs. Such an interpretation is consistent with marine fossils in Bouse Formation from the southern basins (McDougall, 2008; this volume) and depositional environments interpreted to be marine (Buising, 1990). 87/Sr/86Sr values for the Bouse range from 0.7105-0.7115, midway between seawater values and the range of Hualapai Limestone values. Figure 3 shows mixing models using reasonable Sr concentrations and marine versus endogenic end-member isotopic compositions. Mixing curves show that various mixing parameters could reproduce the range of Bouse values (yellow band of figure 3). For springs with Sr concentrations greater than or equal to seawater, and 87Sr/86Sr = 0.735, a mixture of 1-8% endogenic spring water input would produce the slightly elevated values observed for the Bouse. Thus, strontium, carbon and oxygen isotope ranges, and fossil evidence may all be consistent with a marine origin for at least the southern Bouse basins. We caution against interpreting elevated 87Sr/86Sr to negate restricted marine settings, and instead wish to explore mixing models involving endogenic inputs, estuarine conditions, and intermittent marine incursions.

Our conclusions are: 1) Hualapai Limestone was deposited by a groundwater-dominated system (similar to the modern spring/groundwater system in Grand Canyon) such that the carbonates record evolving groundwater flow paths on the Colorado Plateau, with no signal of abrupt arrival of Colorado River water via spill over or karst flow and infiltration of Colorado River water. 2) Models for integration of the different sub-basins of Lake Hualapai at 6 Ma are compatible with our data. 3) Sr, O and C isotope data, and fossil evidence may all be consistent with a marine origin for at least the southern Bouse basins. 4) We favor groundwater sapping as a mechanism for integration of the Colorado River across the Kaibab Uplift and from Lake Hualapai to Upper Bouse. By this mechanism (Pederson, 2001), hydrologic head could have influenced incision and integration by focusing the erosional power of emerging groundwater into surface drainages that were extending across drainage divides. This model explains the observed variations in carbonate composition to be a reflection of different mixing proportions between hydrologic sources in each subregion.

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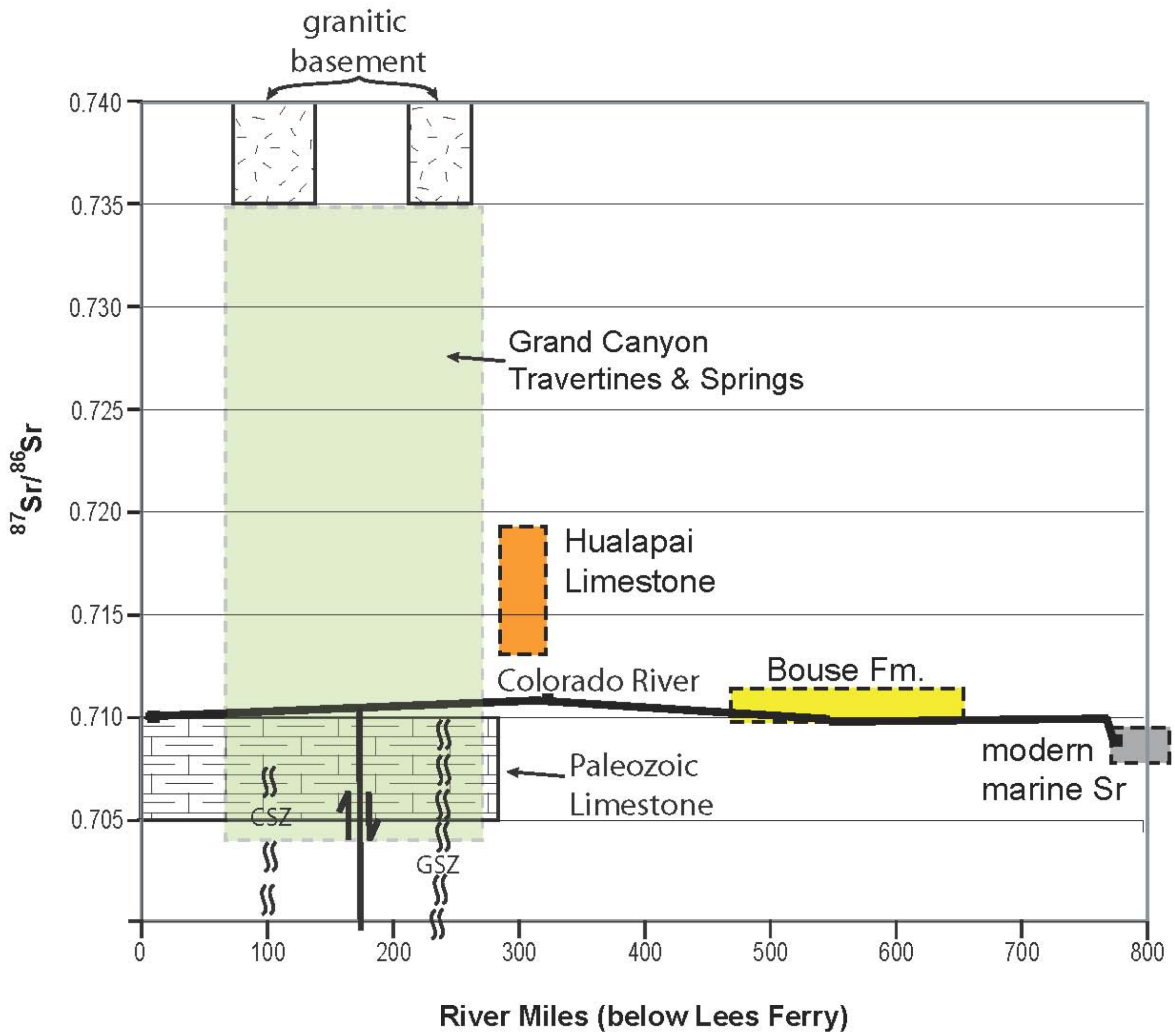

Figure 1. Sr isotope geochemistry for the Grand Canyon region from Lee's Ferry, Arizona, to the Gulf of California (adapted from Crossey and others, 2009). Colorado River composition (heavy black line from Gross and others, 2001) shows downstream increase from river mile 0 to 325, hypothesized here to be due to spring and sidestream inputs of highly radiogenic Sr (green shaded box, data compiled in Crossey and others, 2009). Hualapai and Bouse formations are 6–5.5 Ma carbonates that have elevated Sr relative to marine Sr values, interpreted here to be due to endogenic spring inputs. Basement shear zones separating crustal blocks are: CSZ—Crystal shear zone, GSZ—Gneiss Canyon shear zone; solid fault represents Hurricane-Toroweap active normal faults. 87Sr/86Sr composition ranges for Paleozoic limestones, granitic basement, and modern marine waters are also indicated.

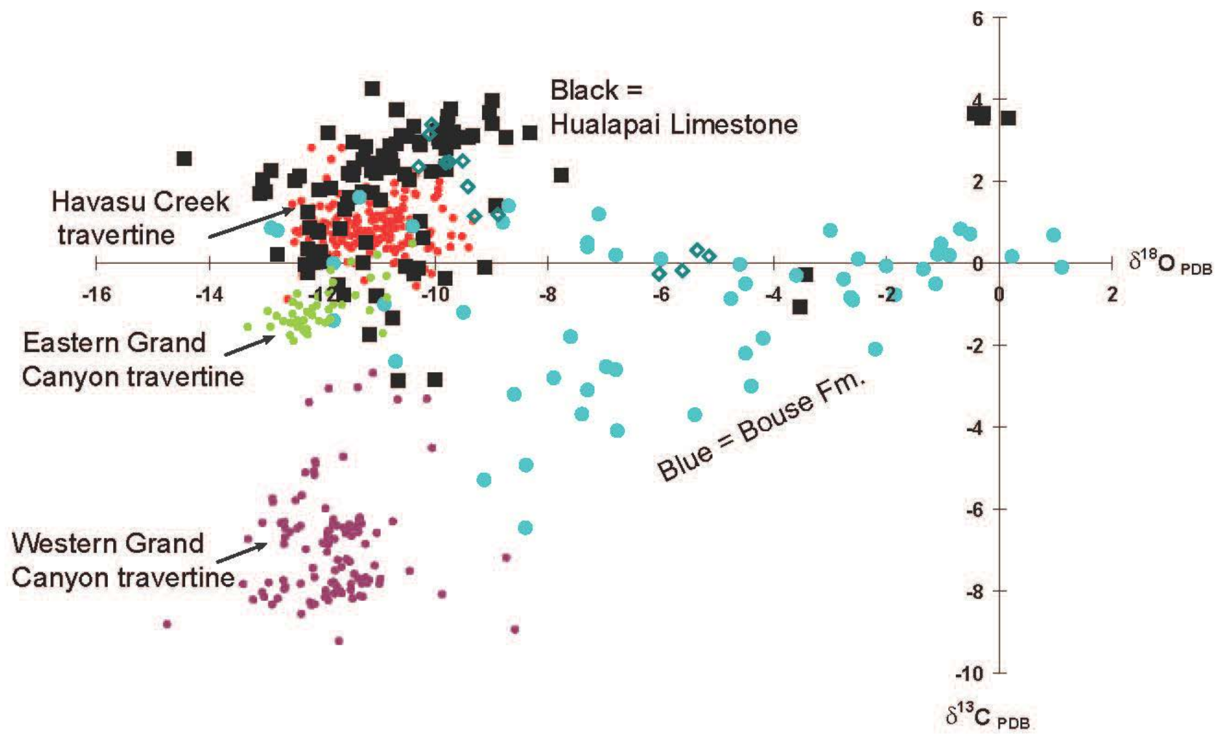

Figure 2. Stable isotopic values of carbon and oxygen for carbonates of the Grand Canyon region. Hualapai Limestone (12-6 Ma) is shown in large black dots (from Lopez Pearce 2010; Faulds and others, 2001). Recent Havasu travertines are shown in red dots (O'Brien and others, 2006). Eastern Grand Canyon travertines (100-350 ka) are shown in green (O'Brien, 2004). Western Grand Canyon travertines (0-380 ka) are shown in purple (Crossey and others, in prep.). Bouse Formation carbonates are shown in blue; open symbols = barnacle shells (Roskowski and others, 2010; Poulson and John, 2003).

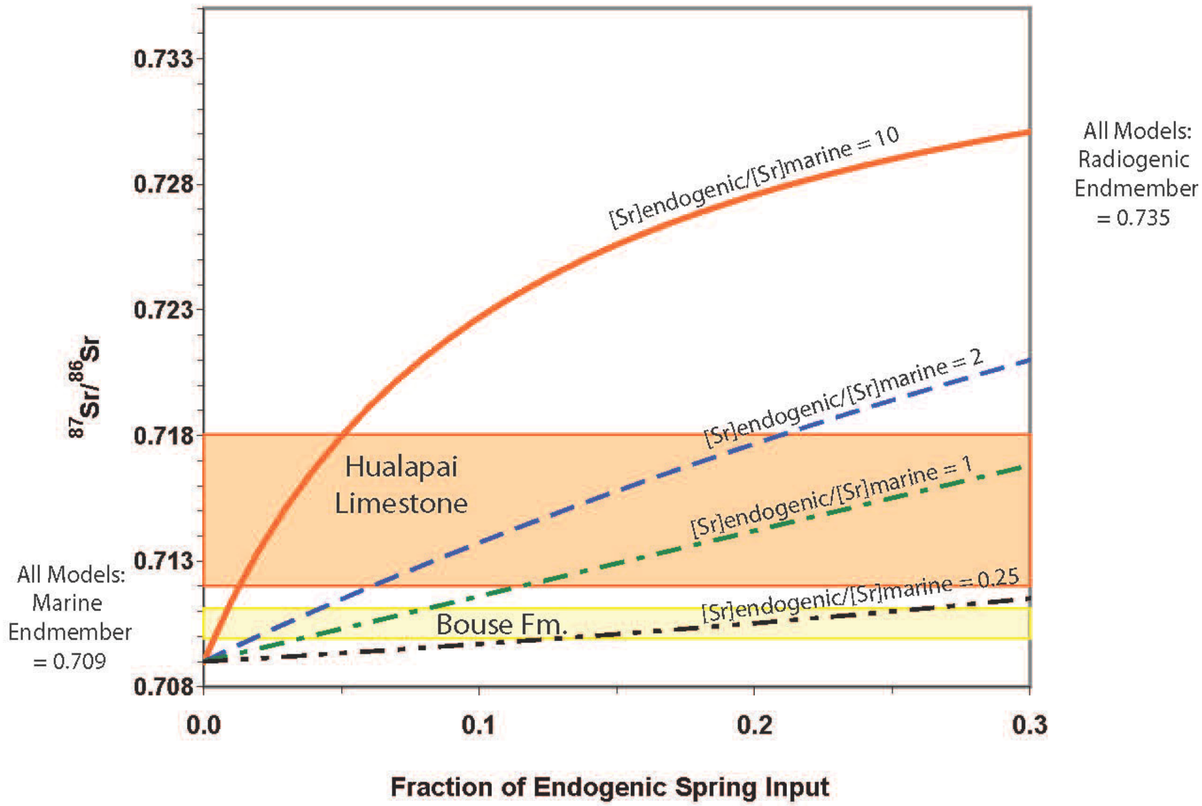

Figure 3. Selected suite of simple binary mixing curves for Sr using an endogenic spring endmember with 87/86 Sr = 0.735 and a marine (estuarine) endmember with 87/86 Sr = 0.709. The curves reflect differing Sr concentrations in the respective endmembers. The orange field encompasses values measured in carbonates from the Hualapai Limestone; the yellow field encompasses values obtained from carbonates of the Bouse Formation. Note that for endogenic Sr concentrations that exceed that of seawater (see top 3 curves), measured Bouse values can be explained by a small component (1- 8 %) of the endogenic water endmember mixing with marine waters.

Incision History of Grand Canyon from Dated Colorado River Gravels

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In spite of 140 years of geologic study, fundamental questions regarding the age and processes that produced the Grand Canyon remain unanswered. The timing and processes of canyon carving are best constrained by the dating of perched river gravels, which directly constrain the past height of the river and thus the subsequent incision rate. In Grand Canyon over 30 incision rates have been calculated by dating basalt flows and travertine deposits which cap and preserve perched gravels and by cosmogenic surface and burial dating of the gravels themselves. This increasingly large data set of incision rates is perhaps one of the most complete records of the spatial and temporal variations in Quaternary bedrock incision on any river and certainly one of the best examples of the effects of faulting on incision. This summary of Grand Canyon's incision rates focuses on the effect of tectonic forcings on canyon incision. Specifically we focus on: 1) how fault-related flexure has affected incision rates in western Grand Canyon, and 2) how large-scale incision rate discrepancies may be due to differential epeirogenic uplift of the Colorado Plateau's edge.

Currently the majority of high-quality gravel-constrained incision rates have been calculated in western Grand Canyon in a reach between river mile[1] (RM) 177 and 246 where 100-630[2] ka basalt flows (Crow and others, 2008 and new unpublished $^{40}Ar/^{39}Ar$ dates) erupted and cascaded into Grand Canyon and flowed more than 120 km downstream. Bedrock incision rates in the reach are predominantly calculated at locations where dated basalts overly gravel-capped bedrock straths using the following equation: $IR = \frac{H + DB}{A + h}$, where IR=incision rate, H=height of the bedrock strath above a 10,000 cfs (283 m^3/s) river level, DB=depth to bedrock below the 10,000 cfs river level, A=age of the datable material (i.e. travertine or basalt), and h=an estimate of the hiatus between deposition of the gravels and emplacement of the datable material.

Active faulting in this reach has produced noticeable offsets of dated basalts (e.g. Karlstrom and others, 2007), is responsible for recent seismicity (e.g. Amoroso and others, 2004), and also affects Quaternary incision rates. Since the amount of folding associated with Laramide compression is hard to quantify (Karlstrom and others , 2007) a younger datum, like Quaternary basalt flows or river profiles, may help to assess the magnitude and geometry of flexures

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[1] This paper cites locations in the canyon in terms of Steven's (1983) river miles, which are measured downstream from Lees Ferry.
[2] While not the focus of this abstract, the new $^{40}Ar/^{39}Ar$ dates also require revisions to the area's volcanic history. Four separate analyses on a duplicate sample from the Spencer Canyon remnant at RM 246, yield a weighted mean age of 567 ± 15 ka, 150 ka less than the previous date of 723 ± 31 ka, which was thought to be the oldest know intra-canyon basalt in Grand Canyon. The best-quality $^{40}Ar/^{39}Ar$ geochronology now constrains intra-canyon basaltic volcanism in Grand Canyon to between about 100 and 630 ka.

associated with reactivated normal faulting. Fault- or uplift-modified total incision rates (TIR) can be calculated as TIR= IR + uplift/downdropping rate (= fault slip rate for many situations) over the same time interval. The concept of dampened incision rates due to relative uplift/subsidence is applicable to a range of tectonic-geomorphic interactions such as fault-dampened incision, salt-induced collapse, differential epeirogenic uplift, and isostatic response.

Eight new unpublished step-heated $^{40}Ar/^{39}Ar$ analyses on groundmass concentrates from Grand Canyon's intra-canyon basalts require revision to 8 previously-published incision rates in western Grand Canyon, which were mostly based on assumed ages (Karlstrom and others, 2007), and give a new incision rate in the immediate hanging-wall of the Toroweap fault. The new data indicate (as shown in figure 1) that, in the block between the Toroweap and Hurricane faults, incision rates increase downstream from 66 m/Ma to ca. 170 m/Ma, with increasing distance from the Toroweap fault. Below the Hurricane fault incision rates vary between 51 and 76 m/Ma. These new data support earlier conclusions that: 1) there is an ~100 m/Ma discrepancy in incision rates between eastern and western Grand Canyon that can be explained by fault dampening of incision rates (Pederson and others, 2002); 2) incision rates in the Uinkaret block, between the Toroweap and the Hurricane faults, are affected by hanging-wall anticline formation associated with the Toroweap fault (Karlstrom and others, 2007); 3) the majority of large-scale fault dampening producing the observed differential incision between eastern and western Grand Canyon is due to slip on the Hurricane fault; and 4) bedrock incision rates were semi-steady from the Quaternary, back through 2-3 Ma speleothem-constrained incision rates (Karlstrom and others, 2008). The new data however require modifications to the geometry of the Quaternary hanging-wall anticline associated with the Toroweap fault, which can be inferred from the rotation of intra-canyon basalt flow bottoms (Crow and others, 2008) and incision rates which are increasingly dampened towards faults due to increased relative down dropping in those areas (Karlstrom and others, 2007). The new data indicate that the folding is shorter in wavelength than previously thought, with incision rates returning to far-field rates about 5 km west of the fault.

In eastern Grand Canyon, gravel-constrained incision rates have only been calculated at a single location near Kwagunt Creek (RM 56), thus there is a 121 river mile (75 km) gap in direct-incision rate data between eastern (RM 56) and western (RM 177-246) Grand Canyon. To test the hypothesis that Quaternary rates are subequal throughout the eastern Grand Canyon (RM 56-177) all the way up to the Toroweap fault (i.e. no foot-wall uplift), which would be expected in the case of epeirogenic uplift of the eastern Grand Canyon, as suggested by Karlstrom and others (2007), we are in the process of calculating incision rates at seven new locations (fig. 2), where dating of gravel-capped bedrock straths is possible by U-series and U-Pb dating of syn-depositional travertine or, in one case, cosmogenic burial dating. At two of the seven locations, multiple bedrock straths have been identified at between 9 and 201 m above river level giving the potential that incision rate variation could be calculated through time. The highest samples at Elves Chasm (RM 116), Hermit Rapid (RM 95), and at RM 273 may be as old as 1-2 Ma, assuming linear extrapolation of Quaternary rates. If the estimated ages are confirmed, they would be the oldest known river gravels in Grand Canyon. As such they would be a critical check on Pliocene speleothem-constrained incision rates (Polyak and others, 2008) which use a controversial water-table lowering model (Karlstrom and others , 2008; Pearthree and others, 2008; Pederson and others, 2008).

As stated above the new travertine-constrained incision rates would also test hypothesis about the epeirogenic uplift of eastern Grand Canyon. Neogene surface uplift of the margins of the Colorado Plateau is consistent with high elevation along the plateau margins, geoid anomalies around the plateau margin (Karlstrom and others, 2008), tomographic and magnetotelluric data, which show low-velocity conductive mantle under plateau margins (Sine and others, 2008; Wannamaker and others , 2008) and numerical modeling of the effect of mantle flow on the Colorado Plateau (Moucha and others, 2008, 2009; van Wijk and others, 2010). Temporal and spatial trends in the isotopic composition and geochronology of Neogene basalts from around the Colorado Plateau also indicates upwelling asthenospheric mantle, which is a plausible driver for Neogene surface uplift of the plateau margins (Crow and others, 2011). The new incision rates will also increase our understanding of Grand Canyon's incision history and constrain numerical models that assess the viability of different Colorado River integration models (Pelletier, 2010).

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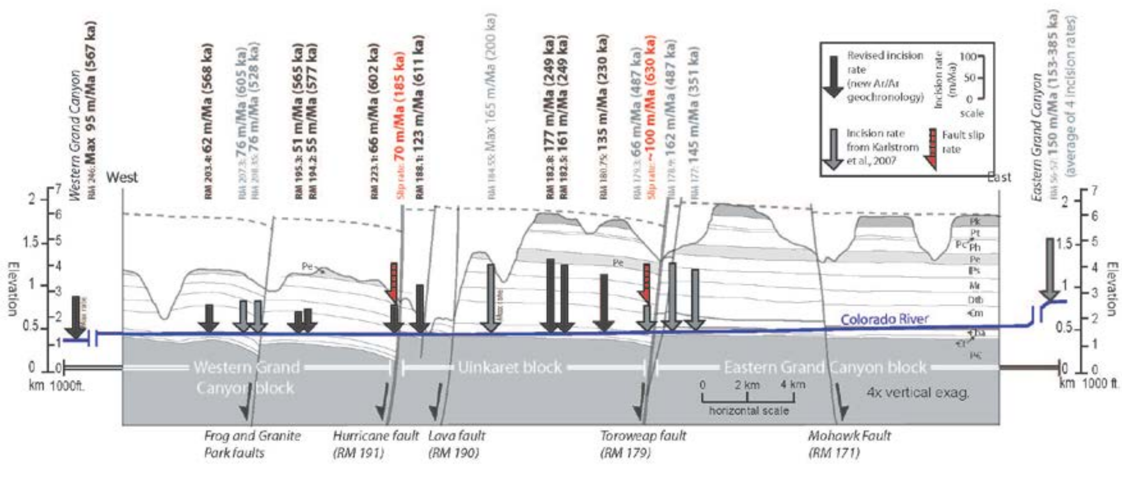

(modified from Karlstrom et al., 2007)

Figure 1. East to west cross-section through the Toroweap and Hurricane faults in western Grand Canyon with the best constrained incision rates, from the 40Ar/39Ar dating of basalts that cap gravel-filled bedrock straths, projected onto the cross-section. Newly refined or previously unpublished rates in black.

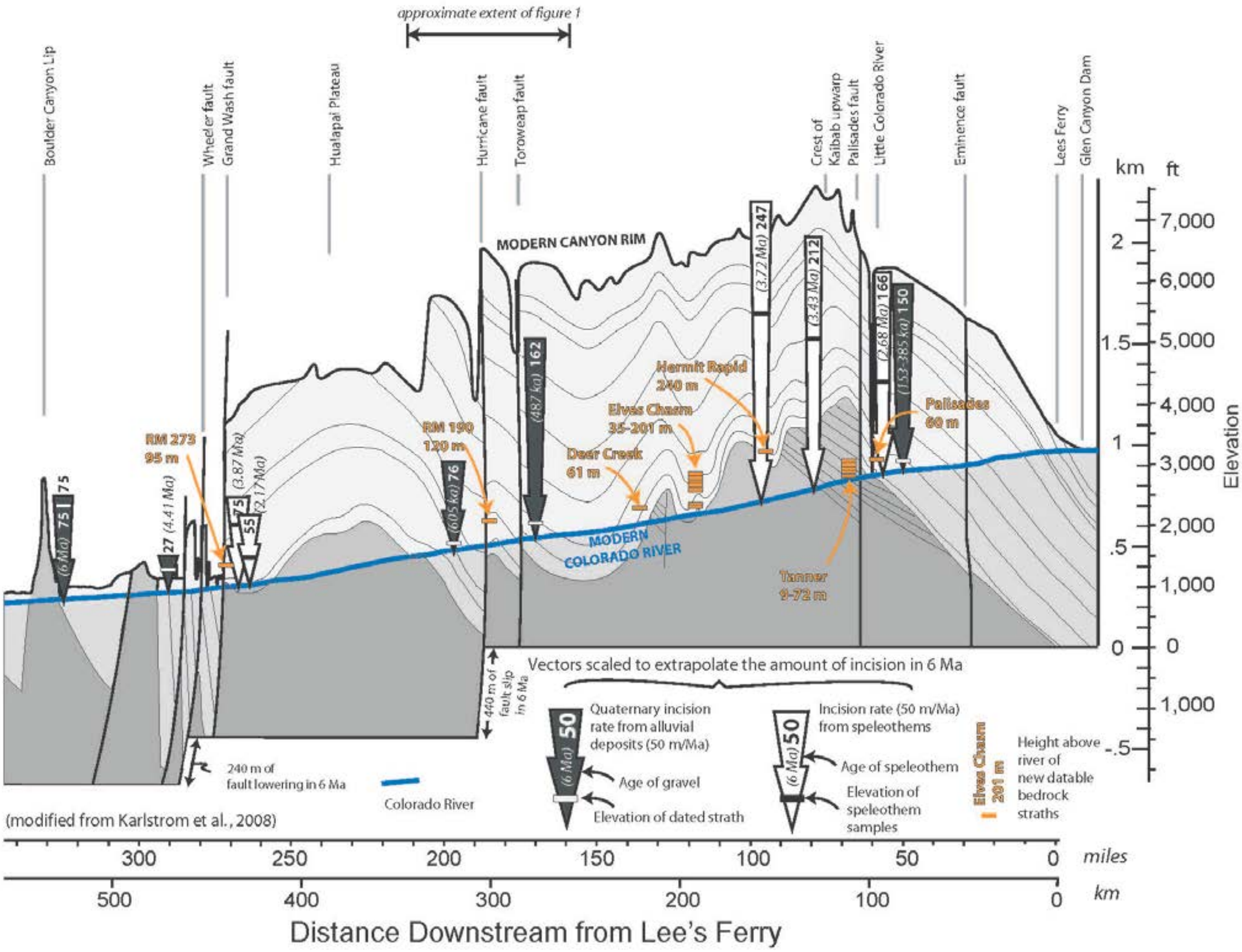

Figure 2. Longitudinal river profile of the Colorado River in Grand Canyon showing a selection of best-constrained incision rates from the dating of alluvial deposits (dark gray arrows from Karlstrom and others, 2007 and 2008) and speleothems (white arrows from Polyak and others, 2008). Heights of dated straths and speleothems are shown by the white and black bars, respectively, for comparison with newly indentified but yet undated travertine cemented river gravels (in orange). Bedrock geology is shown schematically with Proterozoic rocks in dark gray and Paleozoic rocks in light gray.

Differential Incision Rates in the Upper Colorado River System—Implications for Knickpoint Transience

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Longitudinal profiles of rivers in the upper Colorado River system (UCR) reveal large-scale knickpoints and convexities (fig. 1). These features may be a result of numerous processes. For example, hard bedrock, drainage integration events, and/or epeirogenic uplift and faulting all cause or modify relative base-level fall during the evolution of the profile. A major knickpoint at Lee's Ferry separates the Grand Canyon reach of the Colorado from the upper Colorado River profile. Knickpoints above Lee's Ferry include Cataract Canyon on the Colorado and Desolation/Gray Canyons and Canyon of Lodore on the Green River. This paper applies the dating of mainstem river terraces to our understanding of knickpoint formation and evolution. Our overall goal is to understand geomorphic and tectonic processes that have shaped the Colorado River system. Published incision rate data (rate summary displayed in figure 1; data summarized in Aslan and others, 2008) utilize Ar-Ar and K-Ar dates from Miocene to recent basalt flows, deposits of the Lava Creek B ash, travertine U-series dates, cosmogenic surface exposure dates, and cosmogenic burial dates. Ages of overlying datable material provide an estimate of the age of the underlying bedrock strath, whose height is compared to the modern river channel to give incision rates (in m/Ma). We do not use depth to modern bedrock straths in this paper, as those data are generally not available. Hence reported incision rates are considered minima.

Incision rate studies can help evaluate the degree of knickpoint transience, as large differences in incision rates above versus below knickzones necessitate knickpoint migration. As an example, the Gunnison River has incision rates of 150 m/Ma below, 524 m/Ma within, and 95 m/Ma above the Black Canyon knickzone. The regional distribution of incision points of the 640 ka terrace relative to today's river suggests a period of upstream knickpoint propagation in soft sedimentary rock of >150 m/ka from ~1 Ma to 640 ka, which slowed to ~60 m/ka as the knickzone propagated into the basement rocks of the Gunnison uplift. This example shows a 3-orders-of-magnitude larger rate of lateral knickzone propagation relative to vertical incision rates and suggests that knickzones will: a) tend to migrate slower through harder crystalline bedrock, and b) migrate out of the system within a few million years even at the low rates estimated for knickpoints in crystalline rocks.

We report new incision rates from the Colorado and Green Rivers based on burial cosmogenic dating (orange arrows on figure 2). The dates utilize an isochron technique, entailing regression of AMS measurements from 4-6 quartzite pebbles per site to determine an isochron (Balco and Rovey, 2008). Post-burial production (Y-intercept) can be estimated simultaneously with the date (proportional to line slope), allowing less well-shielded deposits to yield viable cosmogenic dates. Comparison of the isochron technique was done with samples related to the Gunnison

River at Bostwick Park. Here we dated a gravel deposit that underlies (by 10 m) Lava Creek B ash (640 ka, Lanphere and others, 2002; Sandoval, 2007). These gravels yield an isochron cosmogenic burial date of 870 +/- 220 ka. The ash is within locally derived canyon fill that post-dated abandonment of the paleo-Bostwick River, perhaps by as much as a few hundred thousand years (Aslan and others, 2008) based on the 870 ka burial age.

Grand and Glen Canyons

In the area of Bullfrog Marina, 180 km upstream from the Lee's Ferry knickzone, we dated a river terrace with a strath 190 m above the pre-Glen Canyon Dam river elevation. Five points yield an isochron defining a cosmogenic burial age of 1.5 +/-0.13 ka. The resulting incision rate is 126 m/Ma (fig. 2). The terrace tread (204 m above the river) was previously dated with a cosmogenic surface date of 479+/-12 ka (Davis and others, 2001). We conclude that the surface date underestimates the terrace age by a factor of ~ 3 due to surface processes such as degradation of surfaces and/or movement of boulders on the terrace surface. A second date was obtained for a terrace 240 km above the knickzone, near Hite, Utah, with a strath 107 m above the river. Its age is 2.9 +0.7/- 0.5 Ma yielding an incision rate of 37 m/Ma (fig. 2). These two ages need to be understood in the context of published rates of 150-175 m/Ma over the last 2-3 Ma below the knickzone (Polyak and others, 2008; Karlstrom and others, 2008), rates on the San Juan River of 100 m/Ma over 1.3 Ma (Wolkowinsky and Granger, 2004), and rates of up to 500 m/Ma over <500 ka dates (Hanks, 2001, Garvin and others, 2005; Cook and others, 2009).

The Bullfrog date is compatible with the age and incision rate from Wolkowinsky and Granger, (2004) and these rates, when compared to higher incision rates below Lee's Ferry, support the idea that the knickpoint is dynamically adjusting and migrating upstream.

High incision rates observed in this region come from both very low and young river terraces and terraces which were analyzed with cosmogenic surface dating and are minimum ages. An increase of incision rate in the last few hundred thousand years to perhaps 300 m/Ma is still allowed by our data, with slower incision through most of the time since Bullfrog and Hite were deposited. Thus we suggest a working model similar to that posed by Cook and others, 2009, such that the transient knickpoint has "diffusely" bypassed Lees Ferry and has been migrating upstream. In their model, the Lee's Ferry knickzone is interpreted to be a transient that has been hung up on moderately hard rocks of the Kaibab Limestone. Their knickzone propagation model suggests that migration is taking place in a diffuse manner via recent and rapid incision in the low gradient region of softer rocks above the knickzone. In the model, incision below Lee's Ferry takes place via upstream migration of the knickpoint through Paleozoic sediments with little or no incision in the reach immediately upstream of the knickpoint until that knickpoint reaches the Mesozoic rock contact. Once the knickpoint reaches the softer rocks upstream of the knickpoint, a rapid but diffusive pulse of incision spreads upstream through the main stem and tributaries. They proposed that the knickzone may have propagated as far upstream as Cataract Canyon, but this does not seem to be supported by the low incision rates at Hite. Due to the difference in incision rate at Hite and the small convexity in the profile below Hite, we suggest that this recent adjustment has not reached Cataract Canyon as suggested, but is possibly immediately below Hite.

It is difficult for an upstream river terrace to be both older and lower in elevation than a terrace downstream of it (fig. 2). A possible reconciliation for the case between Hite and Bullfrog is that one or both dates are inaccurate. As these data are first attempts at burial dating in these terraces, we wish to further test their validity. The isochron for Hite in particular has a lot of scatter which we hope will be rectified with new data being processed at PRIME lab at the time of submission.

However, if we accept the dates, we may be able explain the data using broad tectonic tilting. Only about 0.16° of bedrock tilt is required to restore a 1.5 Ma paleoprofile (with interpolated elevation below Hite) to a profile with the same slope as the modern Colorado River (fig. 2). We suggest that the apparently contradictory elevation and dates for Bullfrog and Hite terraces can be reconciled by a small tectonic tilt due to buoyancy differences observed in the mantle (fig. 2). Given the short wavelength of this feature, current understanding of flexure in the crust may not support this idea, and is only an astonishing possibility.

The overall result of these data leads us to interpret the Lee's Ferry knickzone to be a transient that was set up at the time of integration of the Colorado River system across the Kaibab uplift and Grand Wash cliffs in the last 6 Ma (Karlstrom and others, 2008) and has been responding to both geomorphic and tectonic forcings that include migrating incision waves, differential rock uplift, and isostatic response to denudation.

Upper Colorado River System

A second new set of cosmogenic dates has been obtained from the Green River. A 60 m high strath above the Desolation/Gray knickpoint gives a date of 1.48 ± 0.12 Ma yielding an incision rate of 40 m/Ma over 1.5 Ma. A 120 m terrace within a suite of younger and older terraces near Green River, Wyoming, above the Flaming Gorge/Canyon of Lodore knickzones, yields a date of 1.2 ± 0.3 Ma and an incision rate of 100 m/Ma (fig. 1). Additional dates are needed, but we infer that this differential incision pattern necessitates rapidly migrating knickpoints on the Green River. This may in part be a result of integration of the Green River north of the Uinta Mountains with the drainage south of the Uintas in the Plio-Pleistocene (Hansen, 1986). Current data suggest that the Green and Colorado rivers have been part of an integrated system since sometime after ~8 Ma based on the youngest ash in Miocene basin-fill (Luft, 1985), and well before 1.2 Ma from our data (1.2 Ma is not from the oldest terrace). A regional comparison of the Green and Colorado rivers shows marked contrasts: 1) the Green River has lower gradient, yet has lower discharge, and lower average incision rates over short and long time frames (150 m/Ma for the Colorado and 40-100 m/Ma for the Green (fig. 1). These broad-scale differences seem best explained by differential rock uplift of the Colorado Rockies relative to the Wyoming Rockies and northern Colorado Plateau.

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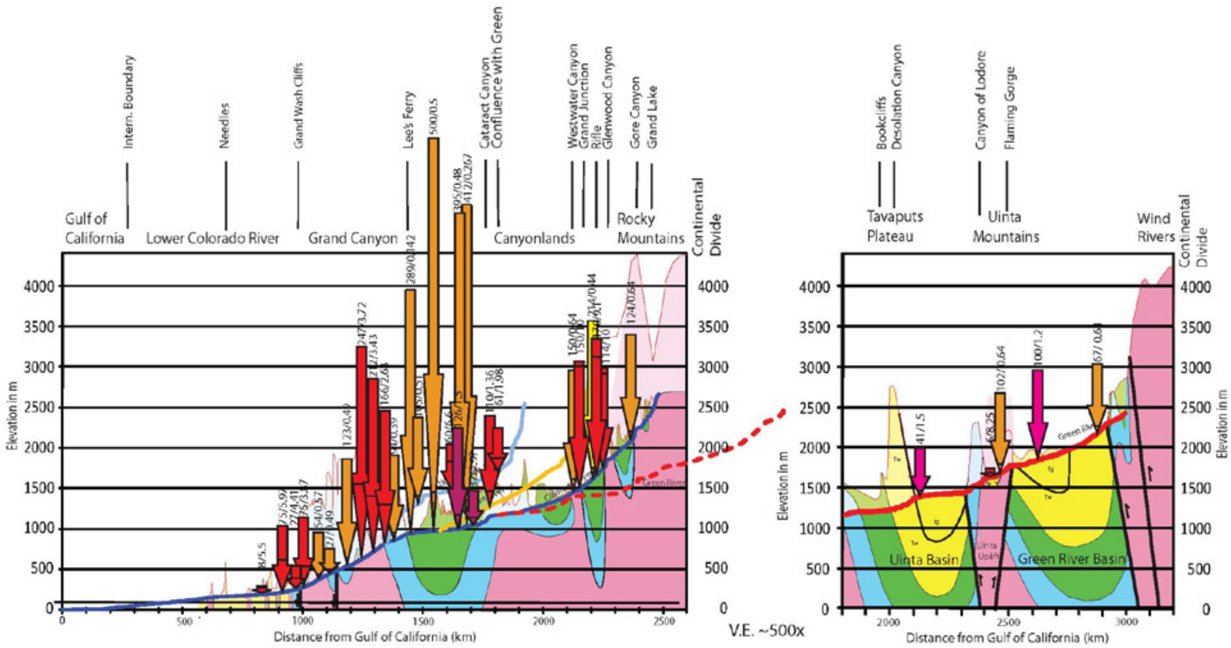

Figure 1. Compilation of incision rate constraints for the upper Colorado River system. Schematic bedrock (V.E. ~500x) shows bedrock below the river and in canyon walls. Incision rates are written as rate/time frame (i.e. m/Ma/Ma) and are scaled to incision rate. Pink and yellow arrows are new cosmogenic data. Red (pink) are long term rates and orange (yellow) rates are short term rates (<1 Ma) River profile is drawn from 1:24,000 USGS topo-maps.

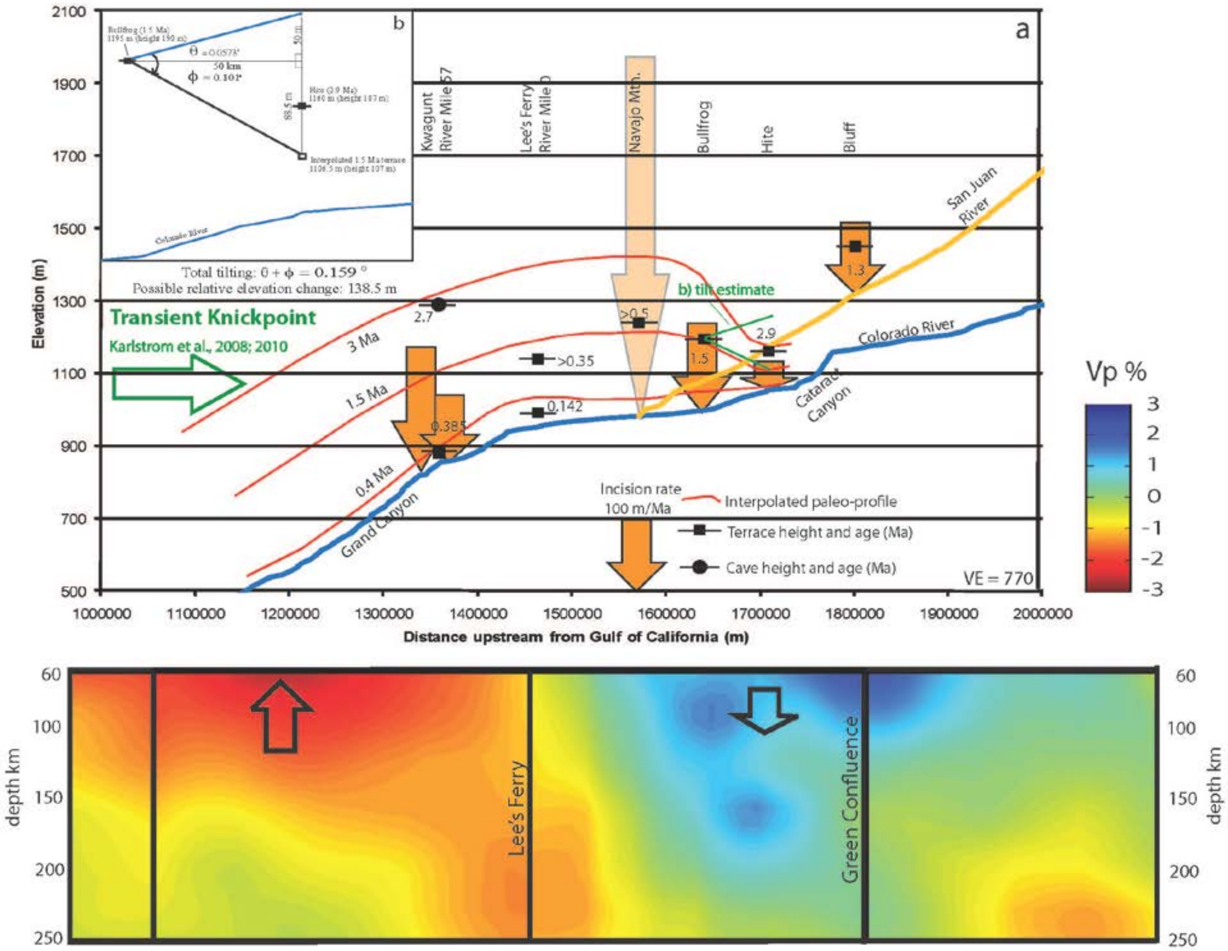

Figure 2. Possible terrace evolution is based on simple interpolation of heights assuming semi-steady rates at each location. More refined analysis will require additional data from different age terraces at a single location to show incision rate changes through time. Tomography is traced along the river corridor.

The Paleogene California River—Evidence of Mojave-Uinta Paleodrainage from U-Pb Ages of Detrital Zircons

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U-Pb age spectra of detrital zircons in samples from the Paleocene-Eocene Colton Formation in the Uinta Basin (northeastern Utah) and the Upper Cretaceous McCoy Mountains Formation of southwestern Arizona are statistically indistinguishable (P=0.23 from K-S analysis). This finding refutes previous inferences that arkosic detritus of the Colton Formation was derived from cratonic basement exposed by Laramide tectonism, and instead establishes the Cordilleran magmatic arc as the primary source (55% of Colton detrital zircon grains are <275 Ma in age and a few are as young as 63-65 Ma). Given the likely existence of a north-south drainage divide in eastern Nevada and the north-northeast direction of Laramide paleoflow in northern Arizona, we infer that a large river system with headwaters in the magmatic arc of the Mojave region flowed ~1000 km northeast to the Uinta basin. The Paleogene California River (Wernicke, 2009) would have been equal in scale but opposite in direction to the modern Green-Colorado River system, and the timing and causes of the mid-Tertiary drainage reversal are important constraints on the tectonic evolution of the Cordillera and the Colorado Plateau.

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Bidahochi Paleogeography and Incision of the Grand Canyon

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The paleogeogaphy of the Miocene Bidahochi Formation exposed in northeastern Arizona and adjacent New Mexico provides critical constraints for the Hopi Lake spillover hypothesis for initial incision of the Grand Canyon (Blackwelder, 1934; Scarborough, 1989, 2001; Meek and Douglass, 2001; Spencer and Pearthree, 2001; Spencer and others, 2001, 2008a, 2008b). Hopi Lake is the name given to the body of water in which the lakebeds of the Bidahochi Formation were deposited.

Figure 1 shows contours on the base of the Bidahochi Formation and the distribution of its lower-middle (lacustrine-volcanic) and upper (fluvial) members. Tuffs in the lower (lacustrine) member (~105 m thick) have yielded Middle Miocene $^{40}Ar/^{39}Ar$ ages of 15.8-13.7 Ma (Dallegge and others, 2003). The middle (volcanic) member (~25 m thick), forming the Hopi Buttes volcanic field, is mapped with the lower member on subregional geologic maps, but has yielded younger Late Miocene $^{40}Ar/^{39}Ar$ and K-Ar ages of 8.5-6.0 Ma (Damon and Spencer, 2001; Dallege and others, 2003). A tuff near the base of the upper (fluvial) member (~60 m thick) has yielded a Late Miocene $^{40}Ar/^{39}Ar$ age of 6.6 Ma (Dallege and others, 2003), compatible with the late Hemphillian faunas (7-5 Ma after the magnetostratigraphy of Lindsay and others, 1994 as recalibrated by Cande and Kent, 1995) collected from the upper member. All available Bidahochi isotopic ages predate integration of the upper and lower Colorado River through the Grand Canyon at ~5.5 Ma (House and others , 2005), but deposition of the fluvial upper member may conceivably have continued into Pliocene time (post-5.3 Ma after Walker and Geissman, 2009). Basalt capping the Bidahochi Formation at Mesa Parada dated ($^{40}Ar/^{39}Ar$) at 2.4 Ma (McIntosh and Cather, 1994) implies, however, that Bidahochi deposition did not persist into Pleistocene time.

The upper fluvial member of the Bidahochi Formation oversteps the older members to rest unconformably on pre-Tertiary strata along the northern, eastern, and southern fringes of the outcrop belt (fig. 1) and the older members are nowhere exposed above an elevation of ~1900 m. The erosion surface below the Bidahochi Formation was incised ~500 m below the base of the Oligocene erg of the Chuska Sandstone overlying an older mid-Tertiary paleosurface that is exposed ~75 km northeast of the Bidahochi Formation (Cather and others, 2008). The configuration of the basal contact below the upper fluvial member shows that the fluvial strata partly occupy paleovalleys of the ancestral Pueblo Colorado Wash, Puerco River, and Carrizo Wash drainages, all now tributary to the Little Colorado River (fig. 1). The upper fluvial member spread over gently sloping (15±3 m/km) pediment-like surfaces carved across interfluves between the principal paleodrainages.

Volcanological analysis of the middle volcanic member in the Hopi Buttes (White, 1990, 1991) indicates that phreatomagmatic eruptions gave rise to multiple diatremes (tephra-filled volcanic necks) and maars, within which subaerial scoria cones were formed locally, but the hydrovolcanism stemmed from contamination of the magmas with water-logged sublacustrine sediment, and not from eruption into lake waters. Water ponded within some maars, but maar rims were nowhere overtopped by lacustrine sediments. Outflow pyroclastic aprons include dry base-surge deposits and subaerial pyroclastic flows that locally were emplaced on muds with desiccation cracks. At the time of Late Miocene eruptions, Hopi Lake in the area of Bidahochi lacustrine deposition was a playa-like body, not very deep and perhaps not perennial (Ort and others, 1998). This interpretation is reinforced by the westward rise in the elevation of the base of the Bidahochi Formation below the lacustrine facies (fig. 1), suggesting that the lake formed within a local depression closed to the west as well as to the east.

Presuming that the water surface of Hopi Lake never reached an elevation more than 100 m above the highest preserved remnants of Bidahochi lacustrine strata, figure 2 depicts the overall configuration of the 2000 m topographic contour on the modern landscape of northern Arizona. The map highlights the difficulties of postulating lake spillover to initiate incision of the Grand Canyon unless the morphology of the landscape was quite different than today, or unless Hopi Lake eventually attained a much greater depth than during the time (8.5-6.0 Ma) of Hopi Buttes eruptions without leaving any preserved record of a lake highstand either within the Bidahochi outcrop belt or elsewhere. Post-spillover erosion of the Coconino and Kaibab Plateaus near the Grand Canyon cannot dispel the conundrum because the plateaus would have stood even higher above Bidahochi exposures before post-spillover degradation than they do today. Perhaps most puzzling is the observation that the Grand Canyon transects the very highest segment of residual highlands extending northwest from the Mogollon Rim, with the lowest modern topographic saddle in pre-Pliocene strata lying due west of Bidahochi outcrops beneath the post-6 Ma San Francisco Peaks volcanic field (fig. 2).

Retention of the spillover model for incision of the Grand Canyon thus seemingly requires the postulate of a tectono-isotatic modification of landscape morphology since the inception of incision. Diverse models for the late Cenozoic paleotopographic evolution of the Grand Canyon region have been proposed (Flowers and others, 2008; Karlstrom and others, 2008; Pelletier, 2010), but none seems at first blush to help resolve the Hopi Lake conundrum, and a paleogeomorphic analysis of the Mogollon Rim south of the Grand Canyon requires no major post-Miocene tectonic modification of landscape morphology in that region (Holm and others, 2001).

The nature of the downstream continuation of the Bidahochi fluvial system remains an open question. Paleodrainages filled by the fluvial upper member had floors with modern slopes of ~0.008 (8 m/km), whereas slopes of the modern Pueblo Colorado and Carrizo valleys are only 0.004-0.005 (4-5 m/km) in the area of Bidahochi exposures and the slope of the modern Puerco valley is ~0.002 (2 m/km), comparable to that of the Little Colorado River between Winslow and Saint Johns south of Bidahochi exposures (fig. 2). The contrast in paleovalley and modern valley slopes suggests that Bidahochi paleodrainages served as piedmont feeders to an ancestral Little Colorado River that formed as fluvial aggradation advanced over Bidahochi lacustrine facies. Modern topography (fig. 2) suggests that the Miocene paleodrainage exited northward into Utah from the Little Colorado lowland but this impression may be misleading. Thermochronology implies that thick successions of Mesozoic strata once covered the Paleozoic strata of the Grand Canyon region (Dumitru and others, 1994), and much of the denudation around the Paria Plateau and nearby modern topographic features probably postdated initial incision of the Grand Canyon.

In sum, Bidahochi paleogeography presents severe challenges to the lake spillover model for origin of the Grand Canyon, as noted previously (Dallegge and others, 2001, 2003). The postulate, for example, that a Hopi Lake highstand once overtopped fluvial lacustrine facies of the Bidahochi Formation seems implausible with present information, yet is seemingly required for retention of the lake spillover model for incision of the Grand Canyon as a viable hypothesis.

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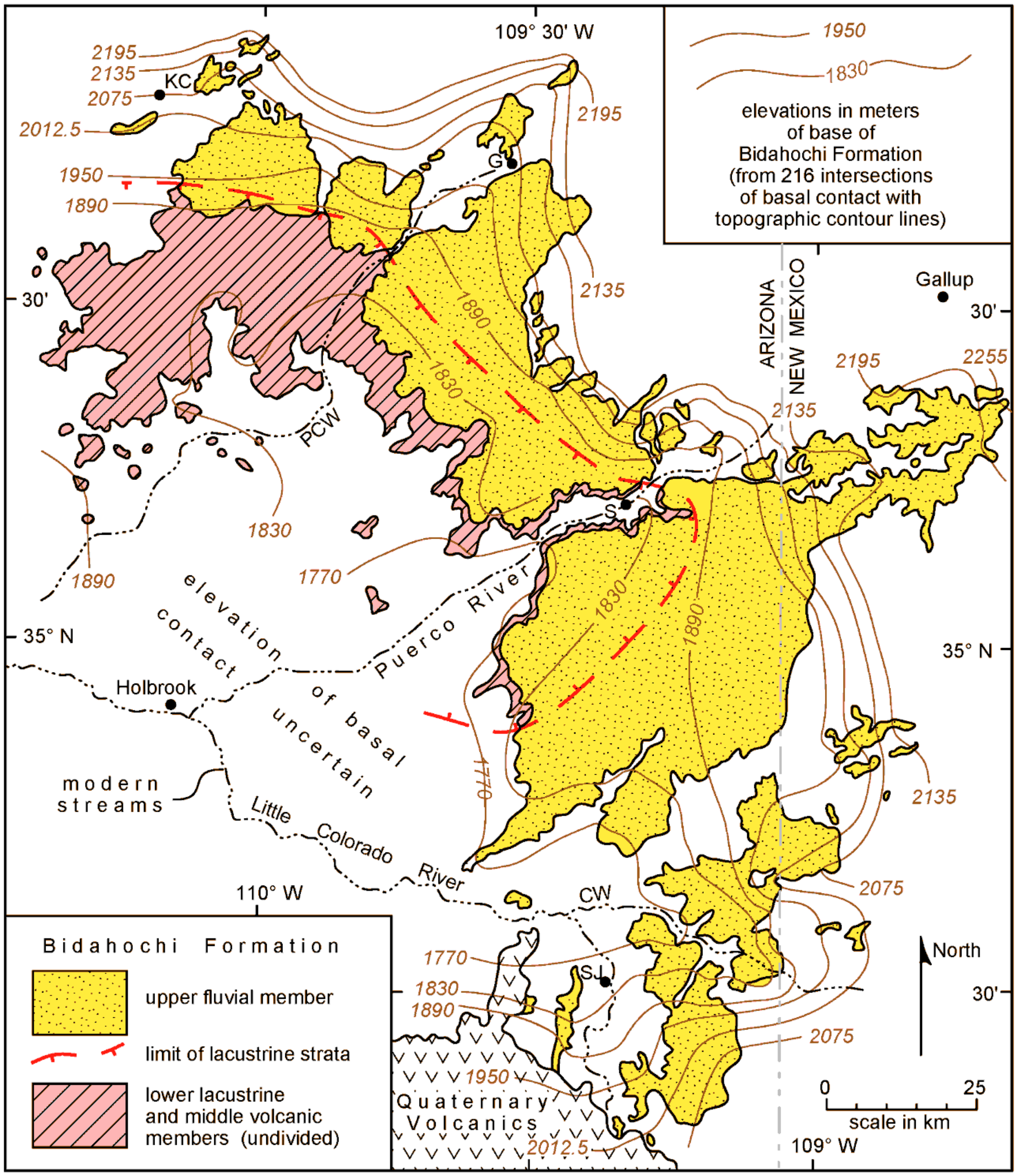

Figure 1. Paleogeographic map of the Miocene Bidahochi Formation, Arizona–New Mexico. Facies and elevations of basal contact updated from Repenning and others (1958) and Love (1989) after Wilson and others (1960), Hackman and Olson (1977), Ulrich and others (1984), Cather and McIntosh (1994), Reynolds (1989), Ort and others (1998), Dallegge and others (2001, 2003), and NMBGR (2003). Abbreviations: CW, Carrizo Wash; G, Ganado; KC, Keams Canyon; PCW, Pueblo Colorado Wash; S, Sanders; SJ, Saint Johns.

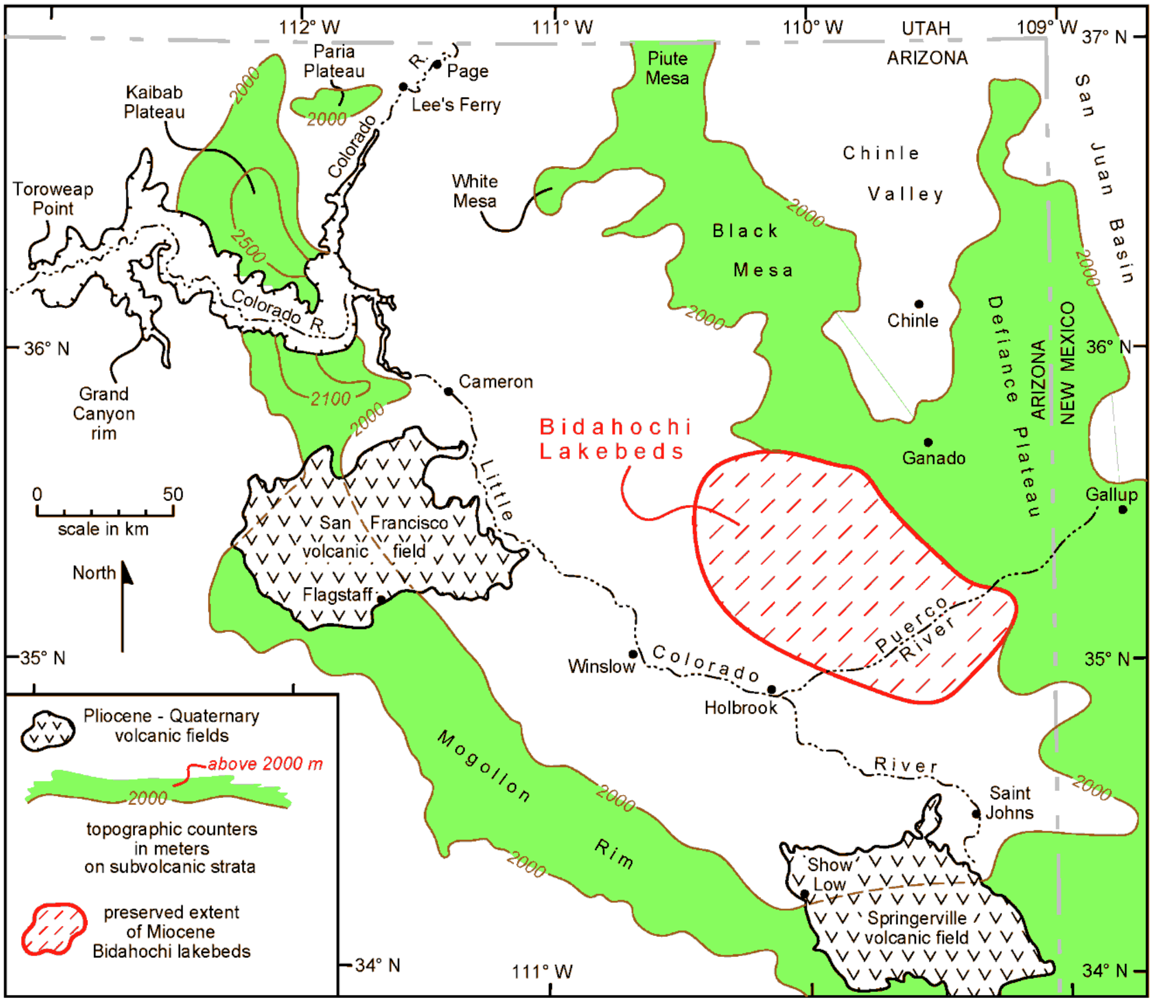

Figure 2. Northeastern Arizona showing extent of Bidahochi lacustrine facies and contours of modern topography on pre-Pliocene sedimentary strata (dashed beneath Pliocene-Quaternary San Francisco and Springerville volcanic fields).

A Preliminary Sediment Budget for the Colorado River

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To better understand the evolution of the Colorado River system, we would like to know more about the timing, rates, and processes by which crust is eroded from the Colorado Plateau, transported down the river channel, and deposited in sedimentary basins along the oblique-divergent plate boundary in the Salton Trough and northern Gulf of California. One way to approach this problem is by calculation of a sediment budget to see if the volume of crust eroded from the source is similar to the volume of material deposited in the sink (e.g., Einsele and others, 1996). While a sediment budget does not directly inform us about driving processes, it can provide a useful framework for more detailed studies, it may reveal gaps in knowledge that help motivate collection of new data, and it can yield insights into regional-scale mass transfer and crustal recycling in response to tectonic and climatic development of the plate boundary and Colorado Plateau. In this abstract I summarize and paraphrase results of a recent study (Dorsey, 2010) to construct a preliminary sediment budget for The Colorado River.

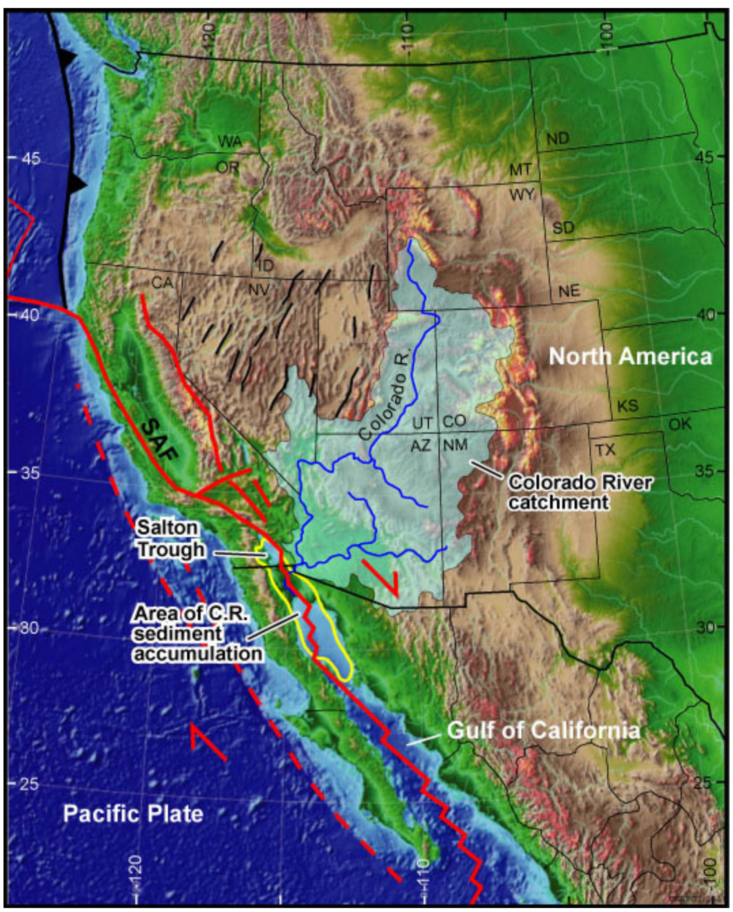

Figure 1. Shaded relief map of western U.S. and NW Mexico (Dorsey, 2010). Shallow bathymetry in the northern Gulf of California reflects large input of sediment from the Colorado River. Area of the Colorado River catchment is 10-15 times larger than the area of sediment accumulation in the basins.

Since ~6-8 Ma, relative plate motion has been focused along the Gulf of California and Salton Trough (Oskin and others, 2001; Oskin and Stock, 2003). This motion has dilated and ruptured the lithosphere to create a series of deep sediment-starved ocean spreading centers in the southern Gulf of California and shallow-marine to nonmarine sediment-filled basins in the northern Gulf and Salton Trough (Fig. 1). The Salton Trough has subsided deeply from late Miocene time to the present in response to regional transtension and extension. The basin has filled with a thick succession of Colorado River sediment, resulting in progradation of the delta to the SE into the northern Gulf.

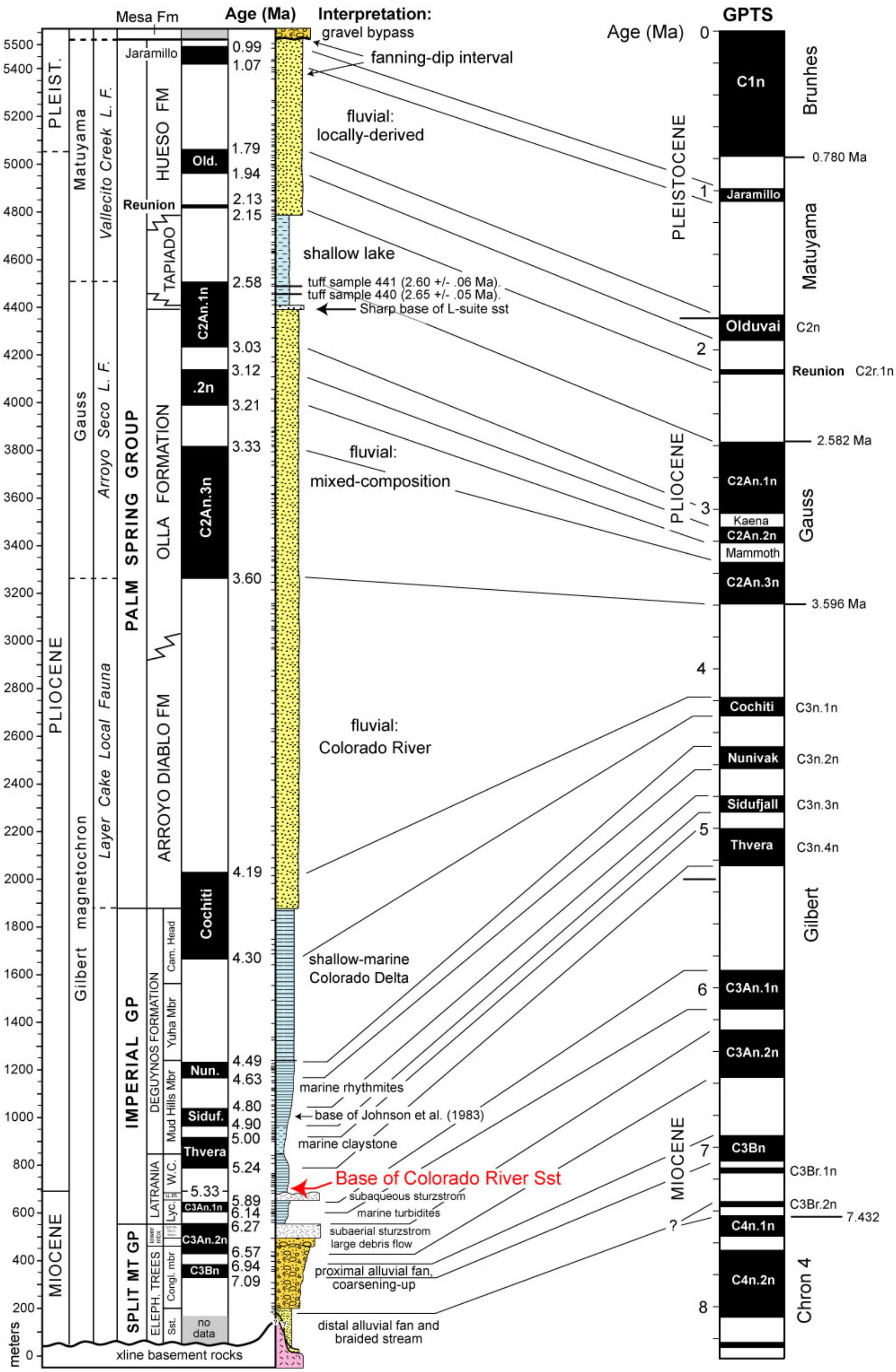

Figure 2. Chronostratigraphy of the Fish Creek – Vallecito basin, western Salton Trough (Dorsey and others, in press).

The first arrival of Colorado River (CR) sand in the Salton Trough is well dated at ~5.3 Ma based on recent detailed studies of lithostratigraphy, paleomagnetism, and micropaleontology (Fig. 2; Dorsey and others, 2007; in press). CR sand is recognized by the presence of abundant fine-grained well rounded quartz grains with syntaxial quartz overgrowths and hematite coatings, chert lithics, and Cretaceous microfossils, all reworked from Mesozoic deposits on the Colorado Plateau.

Colorado River sediment is rapidly buried and metamorphosed in active basins of the Salton Trough and northern Gulf of California, where it is mixed with mantle-derived intrusions and converted to young metamorphic rock (Muffler and White, 1969; Schmitt and Vazquez, 2006). Seismic reflection surveys reveal regionally correlative sequences in fault-bounded basins that drop off quickly to depths of 5-7 km (Pacheco and others, 2006; Aragón-Arreola and Martín-Barajas, 2007; González-Fernández and others, 2005; González-Escobar and others, 2009). Voluminous input of Colorado River sediment exerts a strong though incompletely understood influence on thermal structure, crustal rheology, and rift architecture in this setting. Geophysical evidence suggests that crystalline basement beneath axial sedimentary basins along the plate boundary consists of Late Cenozoic syn-rift sediments that have been rapidly heated and converted to metamorphic rock during basin subsidence and filling (Fuis and others, 1984). Using this model and a range of values for total basin depth, volume of magmatic intrusions, and composition of early rift deposits for six basinal domains, the volume of Colorado River-derived sediment in the basins is bracketed between 2.2 and 3.4 x 10^5 km^3 (Table 1).

TABLE 1. CALCULATION OF SEDIMENT VOLUMES, SALTON TROUGH AND NORTHERN GULF OF CALIFORNIA (from Dorsey, 2010)

Domain #	**1**		**2**		**3**		**4**		**5**		**6**	
Area (km^2)	1,730		3,545		7,300		5,130		17,000		4,750	
	min.	**max.**	**min.**	**max.**	**min.**	**max.**	**min.**	**max.**	**min.**	**max.**	**min.**	**max.**
Sediments*	4	5	4	5	4	4	4	5	4	5	4	5
Metaseds*	6	7	6	7	0	0	4	5	4	5	4	5
Intrusions$^{\Omega}$	0.4	0.1	0.4	0.1	n.a.	n.a.	0.4	0.1	0.4	0.1	0.4	0.1
Non-C.R.$^{\Sigma}$	1	0.1	1	0.1	1	0.1	1	0.1	1	0.1	1	0.1
Volume	11,418	19,376	23,397	39,704	21,900	28,470	27,702	48,222	105,400	159,800	29,450	44,650

TOTAL: **Minimum Volume** = 219,267 km^3 (= **2.2 x 10^5 km^3**); **Maximum Volume** = 340,222 km^3 (= **3.4 x 10^5 km^3**)
* Thickness (km); $^{\Omega}$ Fraction of metasediments volume occupied by intrusions; $^{\Sigma}$Thickness of non-Colorado River sediment (km)

The volume of rock eroded from the Colorado River is estimated using two methods (Dorsey, 2010). First, spatially averaged total erosion of 843 m (Pederson and others, 2002) applied over the Plateau only (3.4 x 10^5 km^2), corrected for the ratio of post-6 Ma to pre-6 Ma erosion (Flowers and others, 2008, their fig. 9), and adding modest inputs from the Virgin and Gila rivers, yields an estimate of ~2.0 x 10^5 km^3. This approach is problematic because it applies information from a small area to the entire Plateau. Second, multiplying pre-dam sediment discharge (1.2-1.5 x 10^8 t/yr; Meade and Parker, 1985) by the time since first arrival of Colorado River sediment in the Salton Trough (5.3 m.y.), and applying an appropriate mass-to-volume conversion, yields an equivalent sediment volume of ~2.5-3.1 x 10^5 km^3 that would have been delivered to the plate boundary basins at early-1900's discharge rates. These are both rather simple, limited, and very preliminary estimates that need to be tested in future work.

Despite existing uncertainties, it appears that the volume of sediment sequestered in deep sedimentary basins along the active plate boundary is roughly similar to the volume of rock that has been eroded from the Colorado River catchment over the past 5-6 m.y. The rate of crustal growth by sediment accumulation in these basins can be expressed as volume (2.2-3.4 x 10^5 km^3) per time (5.3 m.y.) per length along strike of the plate boundary (~500 km), and is roughly 80-130 km^3/m.y./km. This is similar to rates of crustal growth by magmatic accretion documented at subduction-related island arcs (30-200 km^3/m.y./km) and calculated for slow seafloor spreading centers (50-160 km^3/m.y./km). While the contribution of sediment input to crustal growth has been recognized for over 30 years (Moore, 1973; Fuis and others, 1984; Nicolas, 1985), crystalline basement typically is not included in regional sediment budgets. If the crustal model proposed by Fuis and others (1984) applies at other rifted margins, sediment derived from large continental catchments may be partially obscured as it is converted to metamorphic basement in deep rift basins. This process may help explain the origin of transitional crust at some passive continental margins, and could be important at other rifted margins where a large river system is captured following tectonic collapse of a pre-rift orogenic highland.

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One Grand Canyon but Four Mechanisms—Was It Antecedence, Superimposition, Overflow, or Piracy?

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Transverse drainage, drainage that develops across bedrock highlands, only develops via four different mechanisms: antecedence, superimposition, lake-overflow, and piracy. Evidence associated with the four mechanisms was gathered through physical modeling (Douglass and Schmeeckle, 2007) and field research (Douglass and others, 2009) for each mechanism and allows researchers to assess which mechanism is responsible for individual transverse drainage sites with some level of confidence. In this abstract, we compare the expected evidence associated with the four mechanisms to the sedimentary and morphologic evidence relevant to Colorado River's cutting of the Grand Canyon across the Kaibab Plateau.

The antecedence and superimposition mechanisms both require the Colorado River to be older than the most recent exposure of the Kaibab Plateau. An antecedent Colorado River continuously cuts through a rising Kaibab Plateau to maintain grade. The river must drain across the future site of the Kaibab Plateau and therefore predate the structure. A superimposed Colorado River flows across a buried or partially buried Kaibab Plateau. The river would then end up superimposed across the structure after pouring across the Grand Wash Cliffs, ultimately leading to the exhumation of the Kaibab Plateau and the cutting of the Grand Canyon. For this to occur, the Colorado River would need to predate the most recent exposure of the Kaibab Plateau.

The Kaibab Plateau is estimated to have uplifted between 70 and 40 million years ago (Huntoon, 1990). Sediment in the Hualapai basin immediately downstream of the Grand Canyon precludes the rapid arrival of the Colorado River until after 6 Ma (Lucchitta, 1972; Spencer and others, 2001). The oldest sedimentary evidence of the Colorado River, found in Colorado, was deposited roughly 20 to 15 million years ago (Larsen and others, 1975). Pederson (2008) proposed that the Colorado River could have crossed the Kaibab Plateau via superimposition roughly 16 Ma, but then ponded for a prolonged period of time, in a basin between the Kaibab and Shivwits Plateaus, before exiting the Colorado Plateau roughly 6 Ma. However, no Colorado River deposits have been found in the proposed basin between the Kaibab and Shivwits Plateaus.

The presence of the broad southern sweeping course of the Colorado River across the Kaibab Plateau indicates that topography, not bedrock weaknesses, influenced the position of eastern Grand Canyon. The topographic expression of an erosional scarp made up of weakly resistant Moenkopi Formation capped by resistant Shinarump Conglomerate retreating off the southern flank of the Kaibab Plateau controlled the river's position (fig. 1). The Kaibab monocline must already have uplifted for a scarp to retreat away from it, and could not have been buried in sediment. A buried Kaibab Plateau necessary for superimposition requires the covermass to be deposited such that it would allow a through flowing Colorado River to drain down slope to the west, which would bury the retreating scarp and prevent the Colorado River from developing the broad southern curving course. Based on the sedimentary and morphologic evidence, the

Colorado River does not pre-date the most recent exposure of the Kaibab Plateau and, therefore, the evidence does not support a superimposition or antecedence explanation.

Piracy and overflow can explain how the Colorado River crossed an already uplifted and exposed Kaibab Plateau, and both mechanisms require the river to drain to some other location before flowing across the Kaibab Plateau. The upper member of the Bidahochi Formation provides the best evidence currently available to indicate where the Colorado River drained immediately prior to the cutting of Grand Canyon. The Bidahochi deposits widely in northeastern Arizona east of the Little Colorado River, and based on the work of Dallege and others (2001), deposited in a basin that extended westward across the current position of the Little Colorado River to the Kaibab Plateau. Green clays locally found in the Bidahochi Formation's upper member contain the following evidence to indicate the Colorado River drained into this basin prior to cutting the Grand Canyon:

- Fresh water mollusks that lived in a large fresh water lake (Taylor, 1957)
- Fish adapted to fast moving currents including species related to fish currently found in the Colorado River (Uyleno and Miller, 1965)
- >300 m of green lake clays with fossil fish minnows found in a maar crater (Sutton, 1974)

Conformably deposited atop the green clays is a widespread calcium rich sandstone deposit. White (1990) hypothesized this sandstone deposited after a drying phase ended middle member deposition. Based on the evidence listed below, I think the sand rich portion of the upper member deposited around the shore of a large freshwater lake:

- 10x more calcium carbonate in the upper member than the middle and lower members of the Bidahochi Formation, described as a lagoonal deposit (Repenning and others, 1958)
- Calcium rich sand outcrops up to an elevation of 2,225 m where the elevation needed for a lake or river to spill across the Kaibab Plateau is 2,300 m
- The integration of the Bidahochi Basin with the Colorado River that now flows through the Grand Canyon would take place via lake-overflow, drainage reversal, or headward erosion of a bordering drainage divide

Piracy operates four different ways: lateral erosion, headward erosion, sapping, and aggradational spillover. Lateral erosion is likely irrelevant to the formation of Grand Canyon considering the width of the Kaibab Plateau. Headward erosion requires a drainage network to retreat headward into an asymmetrically sloped bedrock structure, steeper on the drainage network side than on the "to be" captured side of the mountain. Spencer and Pearthree (2001) discounted this possibility, based largely on the lack of any relic retreating drainage network on the west side of the Kaibab Plateau. Hill and Ranney (2008) proposed a sapping driven capture of the Colorado River across the Kaibab Plateau. Karst tunnels carved from water by a Colorado River east of the Kaibab Plateau developed in the Redwall Limestone, which eventually collapse along the southern flank of the Kaibab Plateau. The collapse lowers the drainage divide across the Kaibab Plateau, allowing the Colorado River to be re-directed from the Bidahochi Basin into the Muddy Creek basin. However, the presence of the southern bend of the Colorado River across the Kaibab Plateau challenges this possibility. Water flowing through subterranean tunnels will utilize bedrock weaknesses. The resulting collapse would therefore not follow the topographic expression of a circular retreating scarp off the Kaibab Plateau.

The last type of piracy, aggradational, would require a Colorado River to deposit sediment along its channel for a prolonged period of time while draining into the Bidahochi Basin. A river draining into a closed infilling basin would aggrade, but again, the southern bend of the Colorado River poses a problem for this type of piracy. Once the river aggraded to an elevation high enough to pour across the Kaibab Plateau, the river could have followed the path of a circular retreating scarp and formed the *west* portion of the southern bend. The *east* portion of the southern bend also requires a circular retreating scarp, but any scarp would have been buried by an aggrading Colorado River. While each type of piracy could have allowed the Colorado River to pour across an already exposed Kaibab Plateau, all three types struggle to explain the southern bend of the Colorado River across the Kaibab Plateau.

Water pouring from an overspilling lake whose outlet was controlled by the presence of a circular retreating scarp accounts for the broad southern bend of the Colorado River across the Kaibab Plateau. The southern bend has been generated twice on two separate stream table experiments with a 1:60,000 (10x vertical scale exaggeration) scale model of a pre-Grand Canyon landscape, the first filmed by the National Geographic Channel (view here: *http://www.youtube.com/watch?v=SeBPKE5eDU0*) and the second by the History Channel. Water pours from the lake down the pathway of the circular retreating scarp on the east side of the Kaibab Plateau (figs. 2-4). As knickpoints retreats headward and lowers the lake outlet, lake water equal to "the area of the lake x the loss in height to the lake outlet" pours out of the lake, helping to carve the initial 200 m of the Grand Canyon. But as the lake outlet lowers, the circular scarp retreating off the east side of the Kaibab Plateau emerges from the lowering lake water to control the position of the lake outlet. Furthermore, as the lake lowers to the elevation of Cedar Ridge, a broad anticlinal upwarp east of the Kaibab Plateau, the lake separates into a separate northern and southern lake (fig. 4). The northern lake continues to cut forming the initial Marble Canyon and the southern lake continues to cut forming the initial Little Colorado River Gorge. The current confluence of the Colorado and Little Colorado Rivers just offset from the topographic and structural apex of Cedar Ridge is an expected consequence of lake-overflow, but would only happen through pure chance for any form of piracy.

Grand Canyon is one of the world's most striking landforms, but fundamentally, it was carved by the same processes that carve canyons the world over. Grand Canyon was cut by a transverse drainage that only could have formed through antecedence, superimposition, overflow, or piracy. Because the Colorado River is younger than the most recent exposure of the Kaibab Plateau, the available sedimentary and morphologic evidence supports an overflow or piracy explanation, but overflow appears the more likely mechanism.

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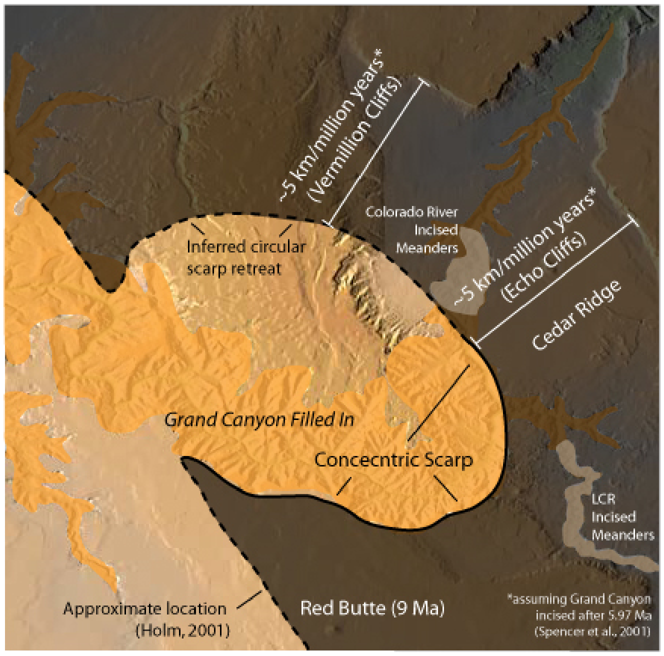

Figure 1. A reconstruction of the pre-eastern Grand Canyon landscape used as the basis for the time series shown in figs. 2-4. The dark grey areas show the proposed location of the Moenkopi-Shinarump scarp immediately prior to Grand Canyon incision. The light orange area is meant to signify that Grand Canyon had not yet been incised and the canyon walls have not yet retreated.

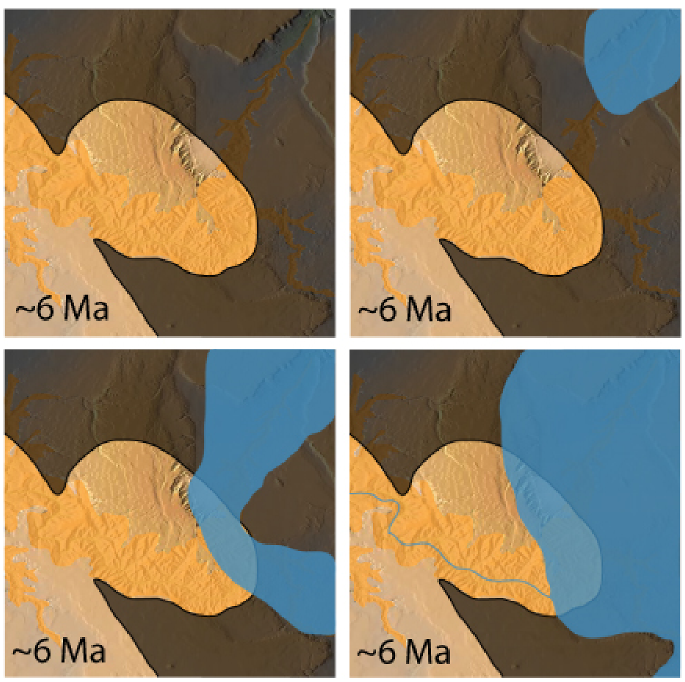

Figure 2. A time series that shows the arrival of the Colorado River and the filling and spilling of Lake Bidahochi across the Kaibab Plateau. The blue areas signify the growing lake over time and the blue line extending away from the lake in the last image represents the water pouring from the lake down the western flank of the Kaibab Plateau.

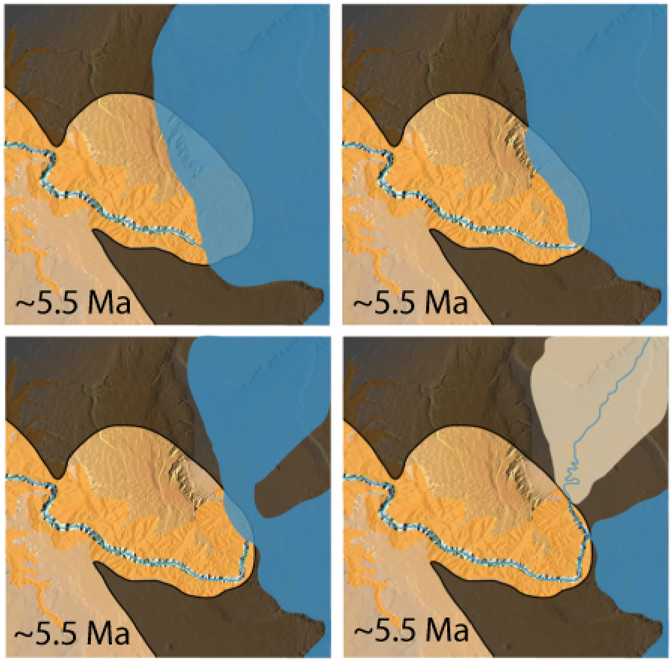

Figure 3. A time series that shows the lowering of Lake Bidahochi whose outlet is controlled by Moenkopi-Shinarump scarp (dark grey areas). As the lake lowers over time from the erosion of the lake outlet, the lake eventually separates into two lakes across Cedar Ridge (labeled in fig. 1). The water pouring from the northern lake is supplied by the Colorado River and the southern lake supplied by the Little Colorado River. The light brown area represents inferred lake deposits exposed once the northern lake was emptied.

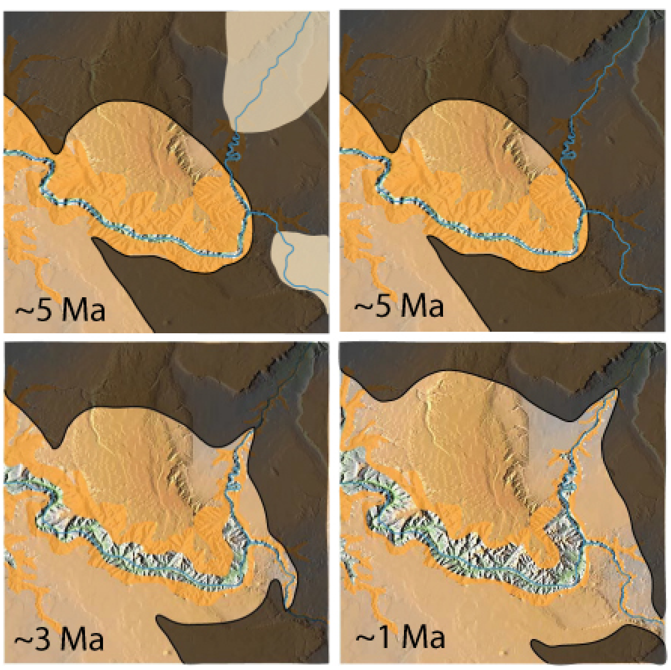

Figure 4. A time series that shows an established confluence of the Colorado and Little Colorado Rivers across Cedar Ridge. Over time, the lake deposits in the area (light brown areas) are eroded, the Moenkopi-Shinarump scarp retreats away from the Kaibab Plateau (dark grey areas), and the retreat of the Grand Canyon walls away from the Colorado River (light orange areas).

Incision History of the Little Colorado River based on K/Ar and Ar/Ar Dating of Basalts and U-series Dating of Travertine in the Springerville Area

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High CO_2 springs and related travertine deposits of the Springerville area of east-central Arizona provide an exceptional field laboratory for understanding travertine-depositing spring systems. U-series dating of travertines provides an opportunity to unravel paleohydrologic and neotectonic histories near the southeastern edge of the Colorado Plateau. A recent interdisciplinary study (Embid, 2010) combines water and gas chemistry data, travertine morphology and geochronology, analysis of geologic structures, basalt geochronology, and river incision studies to formulate an integrative model for both travertine formation and for landscape evolution of this region.

More than 70 individual travertine mounds and large platforms, formed from the coalesced deposits of multiple spring vents, cover a surface area of >33 km^2 near Springerville, Arizona. This area is at the intersection of the southeastern edge of the Colorado Plateau with the Jemez lineament, a northeast-trending zone of volcanic activity over the last 4.5 Ma. Travertine deposits occur in clusters near the Little Colorado River (LCR) and along fault lineaments overlying the Springerville-St. Johns Dome, a faulted asymmetric anticline trapping a large natural CO_2 reservoir. This travertine and CO_2 system is bounded on the west by the Plio-Pleistocene Springerville volcanic field (SPV) which was active until 308 ka and on the east by the late Mio-Pleistocene Red Hill-Quemado volcanic field where volcanic activity continued until as recently as 71 ka.

Modern springs adjacent to the CO_2 field are actively degassing CO_2, have $C_{external}$ values of 50%, concentrations of TDS up to 2538 mg/l, and are currently depositing minor volumes of travertine. $^3He/^4He$ ratios from wells in the CO_2 field and adjoining springs range up to 0.58 R_A, indicating the presence of asthenospheric mantle-derived gases in modern spring waters (up to about 7% of the total helium). To explain the diversity of water chemistry in this small region, we hypothesize that deeply sourced fluids rise along NE- and NW-trending basement-penetrating faults that intersect at the SE end of the dome. These endogenic waters then mix with groundwater producing a complete mixing trend between meteoric and bicarbonate rich, high TDS end members.

Precise new U/Th dates indicate that travertine deposition began >350 ka, overlapping with waning volcanic activity in the Springerville and Red Hill-Quemado volcanic fields, and is still ongoing. Major times of accumulation at 350-300, 280-200, and 100-36 ka are interpreted to represent wetter paleohdrologic intervals. Synchronous outflow occurred from springs at different elevations above the Little Colorado River (from near river level up to 400 m above the river at ca. 200 ka) reflecting an unresolved combination of fluctuations in hydraulic head, gas pressure in the CO_2 reservoir, paleoseismicity, and partitioning dynamics of traps within the stacked CO_2 reservoir system. The life of one major travertine mound system near the LCR that accumulated >20 m of layered travertine has been bracketed between 73 and 48 ka (25 ka). This mound formed from the sustained outflow of CO_2-charged spring waters from a central vent with a deposition rate of 0.94 m/ka.

Dated travertines associated with elevated Little Colorado River gravel terraces and basalts in the Springerville area provide constraints on river incision and landscape denudation. 6 Ma to 2 Ma basalts constrain long-term incision rates of 40-50 m/Ma (fig. 2). Since these basalts lack river gravels, incision rates calculated from their bases should be considered maximum rates. U-series dates on travertine that cements gravels directly above bedrock straths indicate that incision rates increased to ~150 m/Ma near Lyman Lake over the last 280 ka, with rates up to ~300 m/Ma in the last 100 ka. Downstream, near Cameron, Arizona incision rates have been calculated at Black Point and Tappan Spring, where dated basalt overlies a bedrock strath. At Black Point a new Ar/Ar date of 890±170 ka (Hanson, 2010) yields a maximum rate of 183 m/Ma, as gravel occurrences constrain the strath to less than 163 m above river level. The Tappan Springs flow overlies a gravel-capped bedrock strath at 57 m above river level and, based on a K/Ar date of 529±79 (Billingsley, 2001), gives an incision rate of 108 m/Ma.

In comparison with other rates for the Colorado River system (fig. 3), we infer that increased Quaternary rates may reflect migration of the knickpoint presently seen at Lees Ferry on the mainstem Colorado River past the Black Point area in the last ~0.5 Ma, as well as uplift of the southern edge of the Colorado Plateau.

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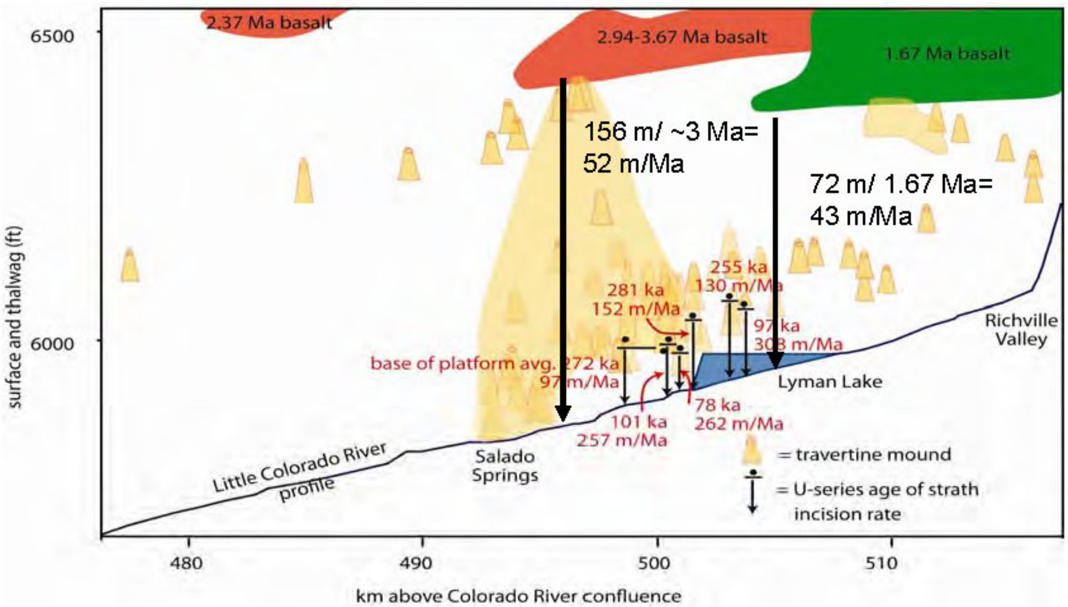

Figure 1. Incision rates in the Springerville area based on basalts and travertines.

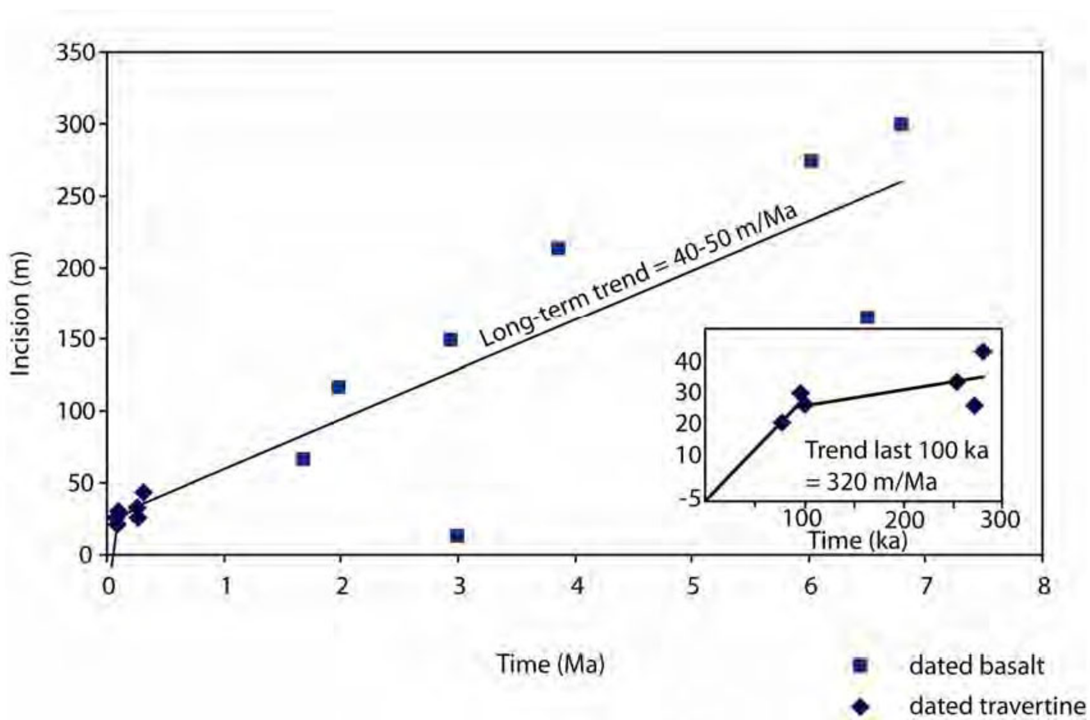

Figure 2. Incision rates through time in the Springerville area based on basalts and travertines.

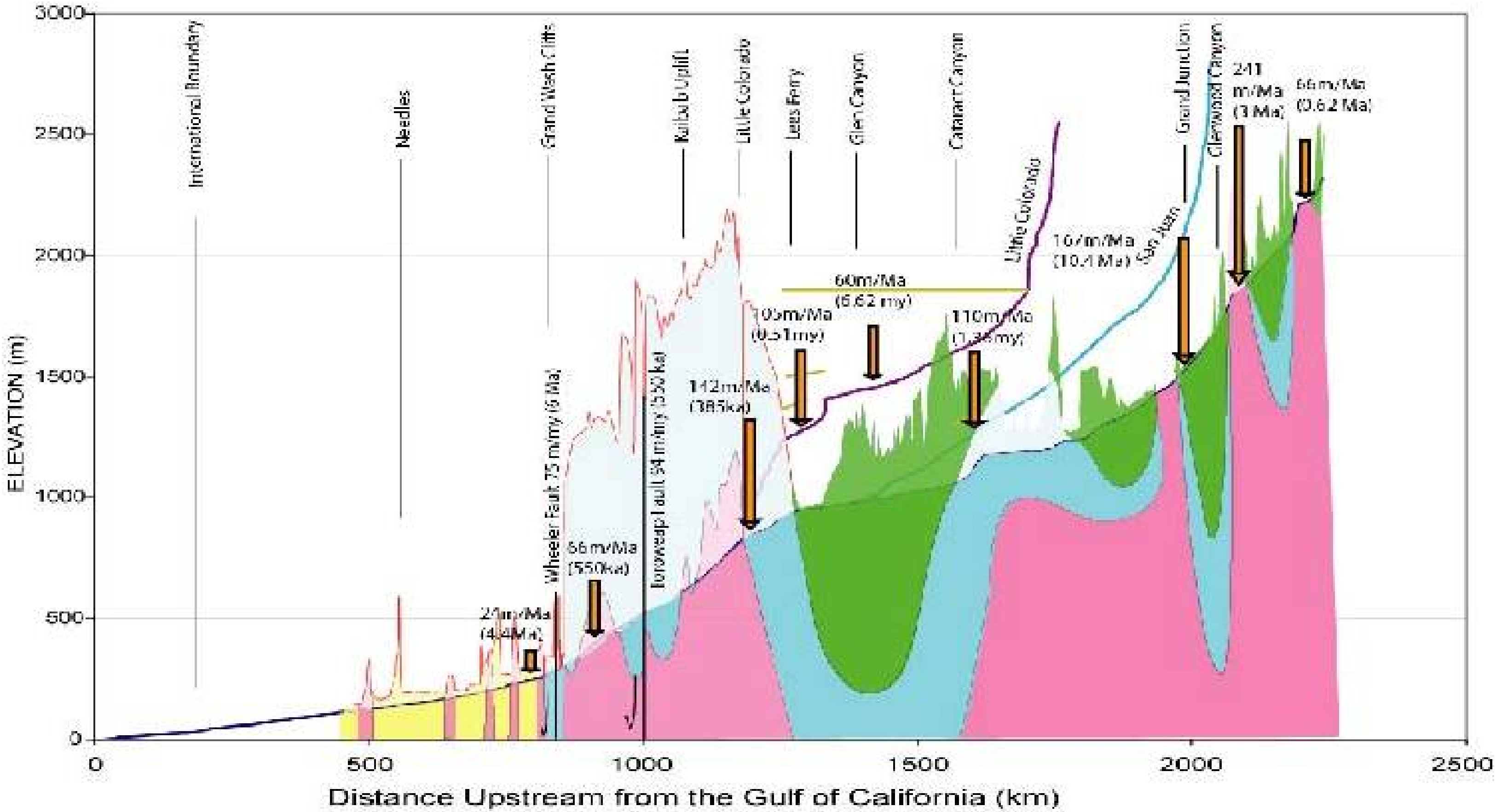

Figure 3. Incision rates on the Little Colorado River in the context of regional Colorado River system rates. Long term rates (in blue) are 40-60 m/Ma over the last 6 Ma; short term rates (in red) are 108-183 m/Ma over the last 300-900 ka.

Miocene-Pliocene Basalt Flows on the East and West Flanks of Wilson Ridge, Arizona, Preserve Multiple Stages in the Depositional History of Adjacent Detrital Wash and Black Canyon Basins, and May Help Constrain Timing of Incision by the Colorado River

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We use eleven new $^{40}Ar/^{39}Ar$ ages and published K/Ar ages for Miocene-Pliocene basalt flows and associated intrusions to help constrain the timing of incision by the Colorado River in the western Lake Mead region, Nevada-Arizona. The basalts range in age from about 5.9 to 4.5 Ma (Reynolds and others, 1986; Feuerbach and others, 1991; this paper), and are preserved at several locations where they overlie or intrude fanglomerate deposits and slope gently away from the east and west flanks of Wilson Ridge, northwest Arizona (fig. 1), graded to a regional base-level. Basalts on the west side include, from north to south (1) Fortification Hill, (2) flow remnants and camptonite dikes (Campbell and Schenk, 1950; Nielson and Nakata, 1994) adjacent to Highway 93 approximately 11 km southeast of Hoover Dam, and (3) Lava Cascade. Basalts on the east side include, from north to south (1) Petroglyph Wash, and (2) isolated flow remnants near the southeast end of Wilson Ridge. The fanglomerate deposits were shed from Wilson Ridge into adjacent Detrital Wash and Black Canyon basins (to the east and west, respectively) prior to incision of the Colorado River, and locally overlie bedrock. The Petroglyph Wash and Highway 93 flows also appear to be interbedded with the fanglomerate, and therefore provide a means for dating it.

Published K-Ar ages for the basalts are: 1) Fortification Hill, 5.89 – 5.42 Ma (Reynolds and others, 1986; Feuerbach and others, 1991), 2) Highway 93 flow, 5.00 ± 0.40 Ma (Reynolds and others, 1986); camptonite dikes, 9.53 – 4.46 Ma (Fleck, 1967; Reynolds and others, 1986), 3) Lava Cascade, 5.16 ± 0.14 Ma and 4.74 ± 0.12 Ma (Feuerbach and others, 1991), and 4) Petroglyph Wash, 4.61 Ma and 5.43 ± 0.16 Ma (Feuerbach and others, 1991).

New $^{40}Ar/^{39}Ar$ ages presented here include eight for the Highway 93 flows and dikes, two for the Petroglyph Wash flows and one for the flow remnants at the southeast end of Wilson Ridge (figs. 1 and 2; tbl. 1). Ages have been recalculated to flux monitor ages consistent with an age of 28.02 Ma for the Fish Canyon Tuff sanidine. At the Highway 93 locality, the large flow bisected by Highway 93 yielded ages of 5.46 ± 0.03 Ma and 5.472 ± 0.026 Ma, whereas the stratigraphically higher flows northeast of the large flow yielded values of 4.852 ± 0.028 and 4.862 ± 0.027 Ma and those to the southeast gave an age of 4.903 ± 0.034 Ma. A complex of camptonite dikes, first studied by Campbell and Schenk (1950), intrudes the fanglomerate at about the same stratigraphic horizon as the large, older Highway 93 flow, and has associated tuff breccia that is inferred to represent a vent area. Three minerals were analyzed from a dike near the inferred vent. $^{40}Ar/^{39}Ar$ age spectra from both kaersutite and plagioclase showed evidence of modest amounts of excess argon, yielding maximum ages of 4.65 ± 0.02 Ma and 4.55 ± 0.19 Ma, respectively. Sanidine, however, gave highly reproducible results by laser fusion analysis at

4.559 ± 0.008 Ma, which is accepted as the age of the dike. The Highway 93 flow and dike ages suggest that the large flow, which overlies fanglomerate at its proximal (east) end and possible axial basin deposits at its distal (west) end, was buried by younger fanglomerate from Wilson Ridge, and subsequently exhumed.

A similar stratigraphic sequence may exist at Petroglyph Wash, where two packages of basalt flows are separated by a boulder conglomerate as much as 10 m thick. A sample from the lower flows, taken just below the contact with the overlying boulder conglomerate yielded an age of 5.743 ± 0.014 Ma, and a second sample taken approximately 2 km to the east yielded an age of 5.711 ± 0.02 Ma. A sample collected from the upper flows was not dated due to groundmass alteration; however, field relationships suggest that the flows overlying the boulder conglomerate were erupted from a diatreme that Feuerbach and others (1991) dated at 4.61 Ma (the exact sample location and detailed analytical data are missing from the published report). If the age of the diatreme is representative of the age of the upper, younger basalt flows, then similar stratigraphic sequences are present at Petroglyph Wash and the Highway 93 locality.

The younger basalt flows at the Highway 93 localities are topographically and stratigraphically higher than the older flows (as opposed to being inset below them), and those at Petroglyph Wash are separated from the lower by fanglomerate, indicating that fanglomerate was still aggrading when the younger flows erupted in both areas. The inferred vent associated with the camptonite dikes at the Highway 93 locality suggests that down cutting had started by the time the dikes were emplaced 4.56 ± 0.01 Ma). Therefore, we infer that down cutting of the fanglomerate in response to lowering base level from incision by the Colorado River occurred sometime between about ~4.9-4.6 Ma in Black Canyon and Detrital Wash basins.

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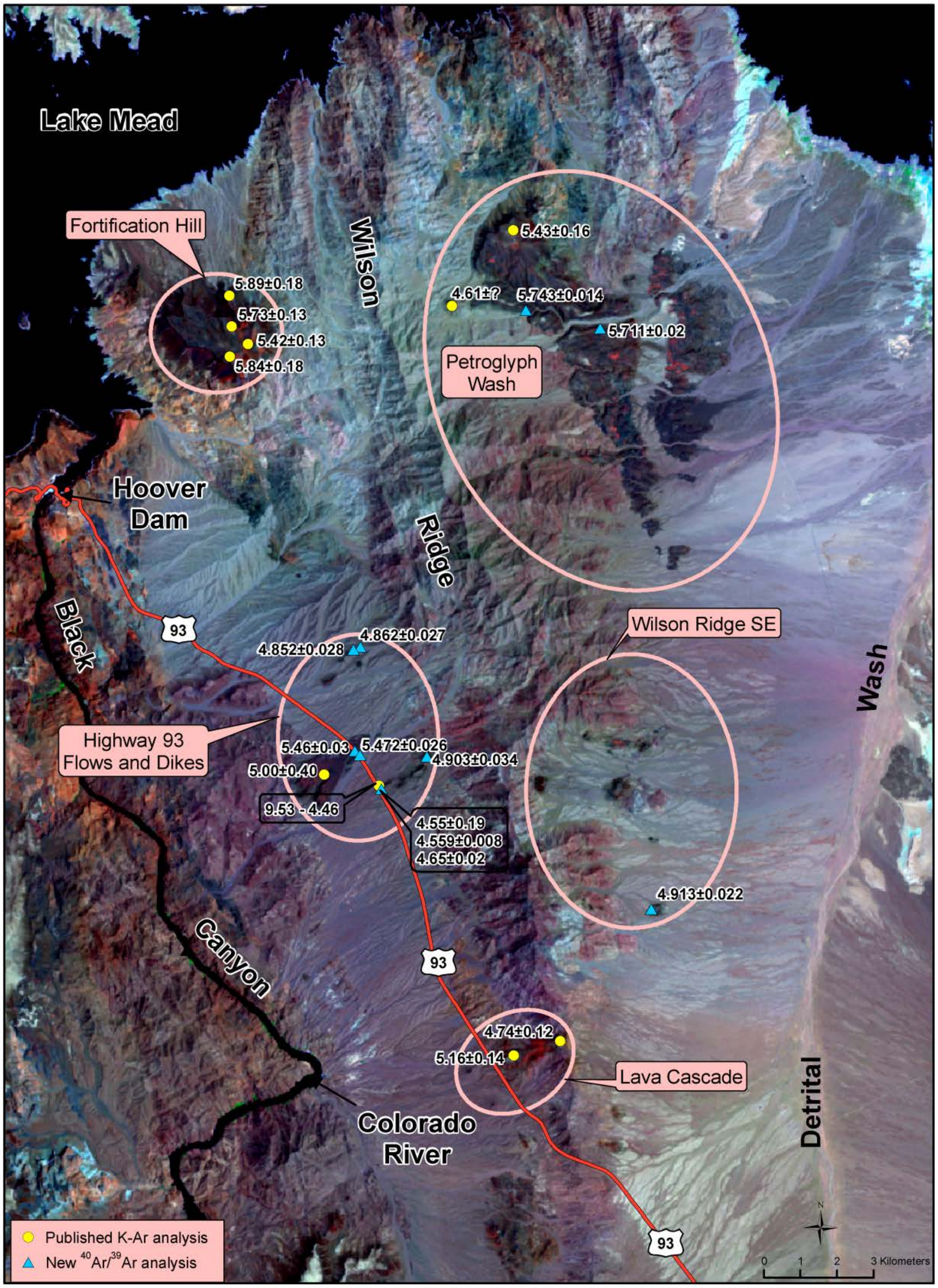

Figure 1. Location of Wilson Ridge flank basalt outcrops and radiometric age analyses. Circled areas delineate outcrops grouped together for the purpose of analysis and discussion. Some published sample coordinates were mislocated by up to 2 km, and have been repositioned to correspond with the analyzed unit.

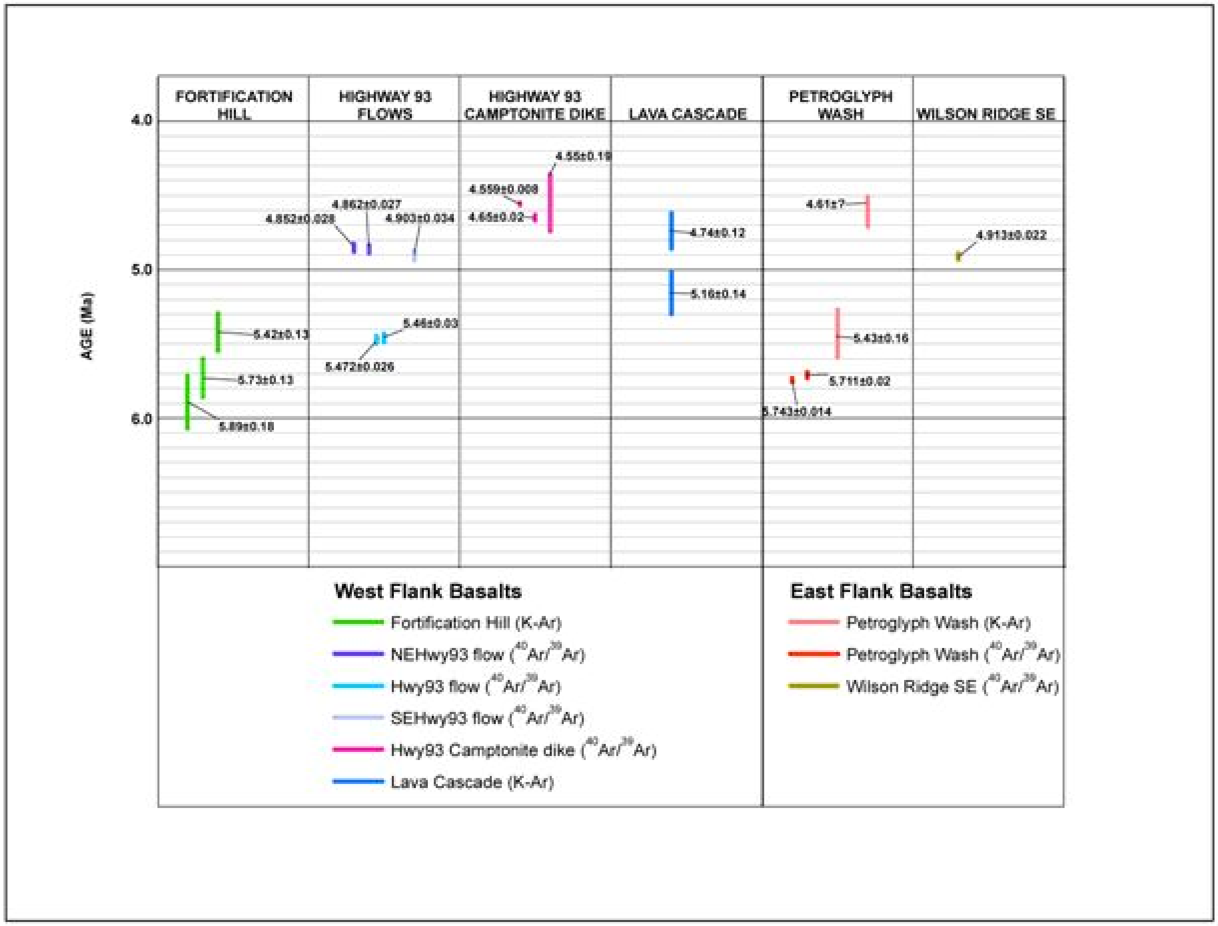

Figure 2. Radiometric age data for Wilson Ridge flank basalts. K-Ar data is from Feuerbach and others, (1991) and $^{40}Ar/^{39}Ar$ data is from U.S. Geological Survey geochronology laboratory, Menlo Park, CA. Figure does not include data from Reynolds and others (1986).

Table 1. New $^{40}Ar/^{39}Ar$ analyses for basaltic rocks of the Wilson Ridge area, Arizona.

Ar-Ar Ages of Basaltic Rocks of the Wilson Ridge Area, Arizona[1]

Sample Number	Material	Location and Unit	Latitude	Longitude	[2]Integrated Age (Ma)	Age Spectrum		Isochron			[5]Indicated age (Ma)	[6]Adjusted age (Ma)
						[3]Plateau Age (Ma)	[7]MSWD	Age (Ma)	[7]MSWD	Intercept (2σ)		
082-19B	Groundmass	Hwy 93; Large Flow	35.94977	-114.64745	5.30 ± 0.03	5.38 ± 0.03	2.43	5.5 ± 0.03	2.5	272.0 ± 5.0	5.38 ± 0.03	5.46 ± 0.03
082-19C	Groundmass	Hwy 93; Eastern Flow	35.94798	-114.62570	4.834 ± 0.034	none		4.82 ± 0.21	4.6	294 ± 42	4.834 ± 0.034	4.903 ± 0.034
082-20A	Groundmass	Hwy 93; Northern Flow	35.97543	-114.64500	4.784 ± 0.028	none		4.9 ± 0.07	2.6	280.7 ± 9.9	4.784 ± 0.028	4.852 ± 0.028
082-20B	Groundmass	Hwy 93; Northern Flow	35.97472	-114.64713	4.794 ± 0.027	none		4.99 ± 0.24	3.3	283 ± 62	4.794 ± 0.027	4.862 ± 0.027
082-21B	Groundmass	Petroglyph Wash Flow	36.05725	-114.59240	5.630 ± 0.03	5.662 ± 0.014	0.47	5.687 ± 0.028	0.35	290 ± 11	5.662 ± 0.014	5.743 ± 0.014
082-21D	Groundmass	Petroglyph Wash Flow	36.05227	-114.57000	5.575 ± 0.031	5.631 ± 0.02	0.37	5.686 ± 0.048	0.16	281 ± 23	5.631 ± 0.02	5.711 ± 0.02
082-21F	Sanidine	Hwy 93; Camptonite Dike	35.94023	-114.63982	4.487 ± 0.026	4.495 ± 0.008	0.397	4.503 ± 0.014	1.10	288.3 ± 7.4	4.495 ± 0.008	4.559 ± 0.008
082-21F	Kaersutite	Hwy 93; Camptonite Dike	35.94023	-114.63982	4.851 ± 0.029	4.58 ± 0.02	0.003	4.58 ± 0.09	17	352 ± 44	4.58 ± 0.02	4.65 ± 0.02
082-21F	Plag	Hwy 93; Camptonite Dike	35.94023	-114.63982	4.725 ± 0.024	4.542 ± 0.013	2.28	4.49 ± 0.19	5.5	504 ± 1200	4.49 ± 0.19	4.55 ± 0.19
H5DW-7	Groundmass	Wilson Ridge SE	35.90931	-114.55944	4.849 ± 0.021	4.844 ± 0.022	1.13	4.8 ± 0.08	0.90	299 ± 12	4.844 ± 0.022	4.913 ± 0.022
H5MtW-15	Groundmass	Hwy 93; Large Flow	35.94870	-114.64677	5.332 ± 0.017	5.395 ± 0.026	1.5	5.38 ± 0.07	5.1	291± 15	5.395 ± 0.026	5.472 ± 0.026

[1]All ages are reported with one-sigma errors.

[2]Integrated age is the age calculated from the sum of all radiogenic ^{40}Ar divided by the sum of all potassium-derived ^{39}Ar in an incremental-heating experiment. "Ma" is the age in millions of years.

[3]An $^{40}Ar/^{39}Ar$ plateau age is the weighted mean age of contiguous steps representing at least 50% of the potassium-derived ^{39}Ar released in an incremental-heating experiment and for which ages are concordant at the 95% level of confidence (Fleck et al., 1977). For $^{40}Ar/^{39}Ar$ laser-fusion data this value is the weighted mean of individual analyses.

[4]The Isochron Age is calculated by weighted-error regression of the $^{40}Ar/^{36}Ar$ and $^{39}Ar/^{36}Ar$ of contiguous gas fractions representing at least 50% of the potassium-derived ^{39}Ar released in an incremental-heating experiment. Ar isotopic ratios are corrected for reactor-derived interfering isotopes.

[5]Indicated Age for each sample represents the age considered the most reliable of those reported here based on interpretation of the $^{40}Ar/^{39}Ar$ data.

[6]Adjusted Age is the Indicated Age recalculated using an age for the flux monitor equivalent to 28.02 Ma on Fish Canyon Tuff sanidine.

[7]MSWD represents Mean-Square of Weighted Deviates, a measure of goodness of fit, comparing the observed scatter to that expected from calculated analytical errors (McIntyre et al., 1966).

Powder Rim Gravel—Deposit of a late Miocene, North-flowing River through the Wyoming-Colorado-Utah Borderland

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A mature, polymict river gravel with north-directed paleocurrents, southerly derived clasts, and a probable late Miocene (~10-5.2Ma) age straddles the Wyoming - Colorado border ~50km north of the east end of the Uinta uplift. The gravel had been correlated with the Bishop Conglomerate (Roehler, 1973; 2004), an enigmatic unit interpreted as a late Laramide, Uinta uplift bajada deposit (Hansen, 1986). New mapping at Powder Rim in southern Wyoming, shows that the river gravel, herein known as the Powder Rim Gravel, is much younger and postdates the mid to late Miocene Browns Park Formation (24-7Ma), a thick-bedded, northwesterly transported eolianite with a thin pluvial basal unit. The age relationship is based on multiple lines of evidence. Most importantly, the Powder Rim Gravel contains large, angular blocks of the Browns Park eolianite along the crest of Powder Rim. A prominent south-side-down normal fault parallels the rim, and juxtaposes 150m of Browns Park eolianite with Eocene Green River Formation. The gravel is preserved only in the footwall (north) where it fills an east-west trending paleocanyon carved into Eocene lake beds. The canyon is thought to have flowed parallel to the fault within the softer footwall strata. The large clasts of Browns Park eolianite within the gravel are interpreted as toppled blocks from the hanging wall in the southern wall of the canyon. Further evidence that the Powder Rim Gravel post-dates Browns Park Formation is in the hanging wall, where no conglomerate or lag occurs within or along the basal contact of the Browns Park. Although basal Brown's Park conglomerates are common in other areas, the others are all very different; generally oligomict, texturally immature, and with clasts derived chiefly from nearby uplifts.

The Powder Rim gravel includes two distinct clast groups. An oligomict suite of friable, sub-angular to rounded, pink quartz sandstone, chert, and limestone clasts decrease in size and abundance from south to north (cobble to pebble, and >60% to <10%). A polymict, resistant suite of well-rounded quartzite>argillite≥chert>>volcanic pebbles increases in relative abundance to the north (<40% to >90%), and does not diminish in size or change shape over the same distance. Powder Rim Gravel clasts contrast with all other Neogene gravels in southwestern Wyoming which contain fairly abundant plutonic and metamorphic clasts. The only exception is a basal Browns Park conglomerate immediately adjacent to the eastern Uinta uplift in northeastern Colorado. This sedimentary-clast-only conglomerate, although a close match to the oligomict fraction of the Powder Rim Gravel, contains no hint of the polymict fraction.

Clasts in the texturally immature fraction of the Powder Rim Gravel are an excellent match for the two resistant rock types that dominate the eastern Uinta uplift; Paleozoic carbonate, and Neoproterozoic through Cambrian quartzite. The decrease in average size and abundance of these clasts to the north is consistent with the north-directed transport for the gravel as indicated

by paleocurrents, and implies a southern Colorado Plateau provenance for the resistant, polymict fraction.

Powder Rim Gravel occurs only in a north-trending, narrow corridor extending from Nipple Rim in Colorado through Powder Rim in Wyoming, where it jogs to the west, and north across Washakie basin to a pinch out near Fort La Clede. The narrow belt is interpreted as an axial river facies. It grades to the west into a belt of piedmont facies conglomerate that cap a series of gently north-sloping upland surfaces flanking the northeastern Uintas. These conglomerates, assigned either as early Miocene basal Browns Park or Oligocene Bishop by previous workers, are reassigned a probable latest Miocene age. The piedmont deposits contain a clast assemblage identical to the oligomict fraction of the Powder Rim Gravel.

The geology at Powder Rim supports an alternative explanation for how the Green River Formation's Atlantic fish fauna went extinct from the northern Plateau since the Eocene. The extinction had been thought to be the result of Pleistocene glaciation (Hansen, 1985), an interpretation challenged by a reevaluation of gravel deposits in southwestern Wyoming (Ferguson, 2008), and genetic biological evidence indicating that Pacific northwest fishes, akin to Bidahochi Formation (northeast Arizona) fossil assemblages, have been in the basin since the late Miocene (Spencer and others, 2008a). Super arid conditions are more likely to have accompanied the extinction (Smith, personal communication), and the impressive, locally 30m thick, cross-stratified sets of eolianite in the Browns Park at Powder Rim offer evidence of extreme aridity in this part of the Plateau in the Neogene. The erg-like deposits of Browns Park Formation at Powder Rim might also provide an explanation for what happened to the Oligocene, southern Colorado Plateau Chuska erg (Cather and others, 2008). That erg may have migrated to the north during the Neogene, and persisted well into the Miocene in Wyoming.

The age of the Browns Park Formation at Powder Rim is crucial since it would provide an older age limit for the Powder Rim Gravel, and help constrain how recently the Plateau drained to the north. The age is unknown, but can be inferred to be 10-7Ma (Izett, 1975; Luft, 1985) based on comparison with similar facies at the top of the formation in its nearby, dated, type area. Undated ash beds occur in the pluvial basal unit at Powder Rim. Detrital geochronologic study (zircon and/or sanidine) of the eolian sandstone at Powder Rim (a subarkose to sublitharenite) might also provide vital information regarding minimum age and additional information regarding provenance which, based on heavy mineral study, have been linked to the San Juan and Elk Mts volcanic fields of west-central and southwestern Colorado (Buffler, 2003).

The Powder Rim Gravel's younger age limit is unconstrained. However, a younger age limit can be inferred based on the conclusion that a major north-flowing river could not have been present in this part of the Plateau during integration of the lower Colorado River at ~5.2Ma (House and others, 2008). This conclusion is based on well-established sediment and water supply arguments. Namely; the essentially unchanged detrital zircon bumpy bar code for early Pliocene and modern Colorado River sands at its delta (Kimbrough and others, this volume), and water budget estimates (Ferguson, 2007; Spencer and others, 2008b) that require a drainage basin similar in size/volume to the current basin during development of the lower Colorado River.

It is becoming increasingly clear that uplift of the southern Colorado Plateau was well under way in the Oligocene or early Miocene (Flowers and others, 2008; Lee and others , this volume), but did not reach the north (southwesternWyoming) until the late Miocene (~8Ma, Naeser, 1984; McMillan and others , 2002) which is now the highest part of the Plateau. The Powder Rim Gravel may have been deposited in the fore of the uplift as it migrated northward into southern Wyoming. An analogy is the northward deflection of early Neogene drainages in southern California (Glazner and Loomis, 1984; Glazner and Schubert, 1985; Loomis and Glazner, 1986) in response to subduction of the East Pacific Rise and Mendocino fracture zone. Northward migration of the north edge of the resultant slab window (Engebretson and others, 1985) since the early Miocene may account for the northward migration of uplift of the Plateau (eg. Dickinson and Snyder, 1979).

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The Colorado River in Glen Canyon during the Pleistocene—Incision Rates, their Uncertainties, and the Possibility of Ancient Impoundments

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In the spirit of this conference to summarize data and results of the ten years since the first incarnation of the Origin and Evolution of the Colorado River System Workshops, this contribution recapitulates data and results closely related to, but more recent than, those presented by Hanks and others (2001) at the first workshop. Hanks and others (2001) determined the incision rate of the Colorado River in its Glen Canyon reach, based on the age and depth of incision of the Cha family of surfaces exposed on the northern flank of Navajo Mountain in southern Utah. The Cha family of surfaces grade smoothly to the edge of Glen Canyon at a nominal elevation of ~1250 m, ~250 m above the pre-dam river level.

The principal theme of Hanks and others (2001), as it is here, is that rapid incision of the Colorado River has occurred in Glen Canyon during the last 0.5 Ma, revealing this mostly steep-walled, ~250-m-deep slot to be a decidedly youthful feature. Unlike many of the major issues concerning the origin and evolution of the Colorado River across the Colorado Plateau, however, this one resides in "close time" and is amenable to further analysis through conventional Quaternary geology and geomorphology, together with modern analytical dating techniques and vastly improved topographic data.

The Cha family of surfaces on the north flank of Navajo Mountain is underlain by alluvial deposits consisting mostly of Navajo sandstone detritus derived from higher elevations of the mountain. This detritus consists of poorly sorted, poorly to moderately consolidated, angular to sub-rounded, boulder-to-pebble-size gravel in a pink sand-and-silt matrix. The angularity and large size of clasts, lack of bedding and sorting, and the sheet-like geometry of these materials indicate deposition by debris-flow. The surfaces that have developed on these materials are smooth and planar, with a few shallow, subdued channels, most of which are truncated by cliff edges. The large (≥1 m) boulders common in the deposits are rare on the surfaces, except at the edges where they are being continuously exposed by lateral erosion, attesting to both the antiquity of the surfaces and the rapidity of lateral erosion. The very smooth, slightly rilled, and more thoroughly vegetated surfaces of the Cha family of pediment remnants easily distinguish them from surrounding bedrock.

The present distribution of Cha family remnants suggest to us that a widespread apron of these deposits covered the northern flank of Navajo Mountain at some earlier time, just as similar (but less dissected) aprons of similar deposits of similar material presently cover flanks of the Henry Mountains. This "earlier time" surely predates the deep dissection of Navajo Mountain, driven by the recent down-cutting of the Colorado River that formed the Glen Canyon visible today.

Hanks and others (2001) estimated the age of the Cha family of surfaces to be ~0.5 Ma on the basis of analyses of pedogenic carbonate, paleomagnetism, and abundances of ^{10}Be and ^{26}Al. Although considerable uncertainty attends this estimate, we doubt that these deposits can be older than a million years, given their normal magnetic polarity and their modest accumulations (Stage III-IV) of pedogenic carbonate. We estimated that the modern slot of Glen Canyon formed in the last 0.5 Ma, at an average rate ~500 m/My.

Garvin and others (2005) extended the analysis of pediment gravels on the north flank of Navajo Mountain with studies at two additional sites (fig. 1). Surface 4103, named for its elevation in feet, is adjacent to the Colorado River in Glen Canyon and a likely remnant of the Cha family of surfaces prior to the last ~250 m of incision of Glen Canyon. This surface is underlain by both locally derived Navajo sandstone detritus and by Colorado River gravels. Oak Island lies within the canyon of the Colorado, ~2 km northeast of 4103, but at an elevation ~90 m beneath 4103. It has existed as an island only since Glen Canyon Dam was closed in the 1960s; its top is covered by Navajo sandstone detritus and Colorado River gravels and is ~170 m above the pre-dam Colorado River.

Garvin and others (2005) assigned an age of ~500±100 ka to the 4103 surface and an age of 250±50 ka for the Oak Island surface on the basis of ^{10}Be abundances, U-series ages of pedogenic carbonates, and the fabric of these carbonates. Thus, the average incision rate of 500 m/My over 500 ka can be refined into two ~250 ka intervals: incision of the Colorado River averaged 400 m/My from 4103 to the Oak Island surface, but had a somewhat faster rate of 700 m/My from the Oak Island surface to the pre-dam level of the Colorado River in the more recent interval. Garvin and others (2005) also found that two nearby tributaries of the Colorado River, Bridge and Oak Creeks, have incised 60 m in the last 100 ka, indicating that their average incision rate of 600 m/My has kept pace with the Colorado River during the past 100 ka.

Cook and others (2009) investigated longitudinal profiles for the major tributaries of the Colorado River in and near its Glen Canyon reach and found that almost all them steepen significantly as they enter the Colorado River, with similar elevation drops (steps) of 150–200 m in the over-steepened segments. These authors attributed this tributary over-steepening to a pulse of rapid incision in all of Glen Canyon during the last half million years. They also found that this pulse of incision has worked its way up Trachyte Creek draining the Henry Mountains in Utah, to a site ~20 km removed from the main stem, incising at the rate of 350–600 m/My during the past 267 ka.

Marchetti and others (2005), using ^{3}He abundances to determine exposure ages of basaltic boulders in two debris-flow fill terraces along the Capitol Reef stretch of the Fremont River north of the Henry Mountains, determined average incision rates of ~400 m/My during the past ~0.2 My. Earlier, in what was principally a study dealing with inheritance problems by amalgamating many clasts into a single exposure-age sample, Repka and others (1997) determined the ages of three strath terraces downstream from Capitol Reef/Waterpocket Fold. Cook and others (2009) estimated incision rates for the Fremont River in this locale to be 300–850 m/My for the last 0.15 My. These Fremont River sites are 150–200 km removed from the Colorado River in Glen Canyon.

Wolfowinsky and Granger (2004) estimated the incision rate of the San Juan River near Bluff, Utah to be 110 m/My, averaged over 1.36 My. This is the "burial age" of this thick set of gravels, and Hanks and Finkel (2005) questioned whether the entire thickness was deposited as a single unit, a common assumption in burial-age analyses. This site is located ~200 km upstream of the mouth of the San Juan River, but not much farther than the Fremont River sites discussed above.

Cook and others (2009) also noted the great spatial variability of published incision rates for the Colorado River and its tributaries in a broad region centered on Glen Canyon, but not confined to it. Together with the transient pulse of incision that washed through Glen Canyon in the last half million years, significant spatial and temporal variability appear to be the rule for Quaternary incision rates for the Colorado River and its tributaries in and around Glen Canyon, not the exception. The challenge here will be to link these diverse estimates together with quantitative models of river evolution that include climatic change and lithologic variations, work begun by Cook and others (2009).

Numerous studies since Hanks and others (2001), then, have reached the same conclusion: Glen Canyon and its tributaries have experienced rapid, though variable, incision rates since ~0.5 Ma of hundreds of m/Ma or greater. Most of these incision rates, however, come with significant uncertainties that are mostly underestimated and often left unspoken. Although the analytical uncertainties of various dating techniques are known, they are mostly dwarfed by the uncertainties of what any sample has experienced in its space-time trajectory to the sample location. Important distinctions exist between sample ages, surface ages, and depositional ages. Model calculations to determine surface or depositional ages from sample ages often involve assumptions that are difficult, if not impossible to verify (Hanks and Finkel, 2005). Any incision rate involving a depth of incision of the order of the (unknown) river-channel dimensions is subject to large uncertainties, of the order of the incision rate itself (Hanks and Webb, 2004). Moreover, such incision rates must distinguish between bedrock incision and debris-fill incision in the case of alluvial rivers, which the Colorado River has become in Glen Canyon and much of Grand Canyon (Hanks and Webb, 2004, 2006). Incision rates involving greater incision depths and correspondingly greater intervals of time can be more accurately determined, but at the price of averaging over long periods of time that may obscure or hide significant transient signals.

The importance of uncertainties relates to the admissibility of viable alternative models, hazard-and-risk-analysis newspeak for what G. K. Gilbert, a century earlier, called multiple working hypotheses. Any physically plausible model that fits the data within their uncertainties is a viable alternative model, and the idea is to retain them all until the observations themselves exclude one or more of them. At the present time, several such models exist for the evolution of the Colorado River across the Colorado Plateau during the past 5–6 My—and also for the past 0.5–0.6 My. The significance of a set of viable alternative models lies in their potential to define further research at critical sites that may discriminate among the differing models.

The low-gradient Glen Canyon is sandwiched between two steep reaches of the Colorado River, Cataract Canyon at the upstream end and Marble Canyon at the downstream; both are the falling limbs of significant convexities in the river profile. Neither convexity seems to be the consequence of retreating (upstream propagating) knickpoints. The Cataract Canyon convexity is underlain by up to 80 m of alluvial fill, on which rests the steepest reach-averaged gradient of the Colorado River (Webb and others, 2004; Hanks and Webb, 2006). This convexity, we believe, is the consequence of latest-Quaternary aggradation that the enervated Holocene river is unable to keep up with, just as it cannot in various parts of the Grand Canyon (Hanks and Webb, 2006). The Lees Ferry convexity is very different, underlain by the hard, less-erodible upper Paleozoic rocks of the Grand Canyon section. Cook and others (2009) suggest that the ~0.5 Ma Glen Canyon incision pulse was caused by a comparable incision event propagating upstream from the eastern Grand Canyon, and evidence for this event may exist in the gravels atop Johnson Point near Lee's Ferry (Davis and others, 2001; Pederson and others, this volume). Cook and others (2009) believe that the Lee's Ferry convexity is a consequence of lithology, not of a propagating knickpoint.

Quartzite river gravels crop out from one end of Glen Canyon (Wahweap Basin) to the other (Hite) at or near the top of Glen Canyon, with a nominal elevation of ~1,250 m asl as assigned by Hanks and others (2001), as well as at lower elevations within the canyon (Oak Island, for example). (Most, but not all of the lower elevations are now covered by Lake Powell at a full-pool elevation of 1,128 m.) The gravel deposits in the Wahweap and Bullfrog Basins (fig. 1) are huge, extending many miles back from the river, although base-level fall since the recent incision pulse has allowed the modern tributaries to excavate much of the gravel that previously existed. Wahweap and Bullfrog Creeks carry predominantly quartzite gravels from the late-Cretaceous Canaan Peak Formation to the north. These gravels are distinguished from "Colorado River" quartzites only by the presence of litharenites first described by Goldstrand (1992). Wahweap Basin is just upstream of the Echo Cliffs, and Bullfrog Basin is just upstream from the Waterpocket Fold. Prior to the existence of a through-going Colorado River, both of these monoclines could have impounded large lakes behind them, and basal members of the river-gravel deposits in these basins could be quite old.

To build on the interest at the Workshop in "spill-over" mechanisms, with or without subterranean flow, as significant players in the integration of the Colorado River system (Douglass, this volume; House and others, this volume; Hill and others, this volume), figure 1 illustrates the possibility of ancient ponds or lakes in the Glen Canyon area prior to the incision of Glen Canyon beginning ~0.5 Ma. Figure 1 shows the Glen Canyon reach of the Colorado River with its significant tributaries; in yellow are elevations between the 1,250 m contour, the nominal top of Glen Canyon, and the (arbitrarily chosen) 1,300 m contour above it. The Wahweap and Bullfrog Basins, as defined by the yellow elevation band, suggest ancient impoundments of water, consistent with the gravel deposits within them.

We believe many other gravel deposits are yet to be found adjacent to or within Glen Canyon. The average gradient of the top of Glen Canyon, to the extent it is definable, is very low, much like the pre-dam river gradient, ~0.2 m/km until it approaches Cataract Canyon. The 1,250 and

1,300 m contours extend far up the Green and Colorado River bottoms, beyond the margins of figure 1 to the Price River in Gray Canyon on the Green River and almost to the Colorado-Utah boundary on the Colorado River; they suggest a level of stability for long periods of time prior to the recent incision of Glen Canyon.

The 1,300-m contour is close to the top of the Echo Cliffs where the Colorado River exits Glen Canyon, and it would have formed a closed basin on the Marble Platform downstream had the Grand Canyon not been there to drain it. Also of interest is the wide expanse of yellow in the basin traversed by the San Juan River just before it enters Lake Powell. A larger version of figure 1 extending farther upstream is in the set of maps prepared for the conference by J.L. Blair (*http://sites.google.com/site/crevolution2/home/files*).

One of many curiosities of the Colorado River, as it traverses the Colorado Plateau for a thousand miles or more, is that it generally flows up the dip of the strata that underlie it. Even more enigmatic are the processes by which the river crossed various monoclines, often with considerable topography to (seemingly) ascend and then to incise. The most significant of these is the Kaibab Upwarp, but the Echo Cliffs and Waterpocket Fold present considerable topography–and comparable problems–further upstream in Glen Canyon. As indicated in figure 1, both the Echo Cliffs and the Waterpocket Fold may have served as impoundment barriers (or "dams") prior to the incision of Glen Canyon beginning at ~0.5 Ma. Perhaps the Colorado River "learned" what it needed to know to traverse the Kaibab Upwarp by first practicing on the Echo Cliffs, Waterpocket Fold, and other monoclines further upstream.

Evidence for significant ponding of meteoric water on the Colorado Plateau long before integration of the Colorado River into something like its present course has existed for decades. On the Hualapai Plateau in the southwest corner of the Colorado Plateau, Young (1999) noted the 60 m of white limestone logged in four early Santa Fe Railroad wells on the upstream (upgradient) side of a projected monocline axis where it intersects the deep paleovalley that underlies the town of Peach Springs, Ariz. This limestone overlies (or possibly interfingers with) coarse arkosic sediments now mapped as the Music Mountain formation. A 14-m-thick white lacustrine limestone, also part of the Music Mountain Formation, lies on the upgradient side of the Meriwhitica monocline that intersects the Milkweed Canyon paleovalley (Young, 1999). Northeast of the Hualapai Plateau, near Long Point on the Coconino Plateau, tens of meters of early Eocene fossiliferous limestones in the Music Mountain Formation (Young and Hartman, this volume) lie upstream from the southeastward trending Supai monocline on a stripped Moenkopi-Kaibab paleosurface. The Long Point lacustrine limestones reveal that all Mesozoic and younger strata here were removed by early Eocene. Their preservation for ~50 My on the Kaibab limestone is ample testimony to its toughness and inerodibility. In a more general manner, Young (2008) discusses the nature of such ponds on the Coconino/Kanab Plateaus in terms of another pond, in which the Hualapai limestone was deposited at the downstream end of an incipient Grand Canyon.

Finding and mapping the geologic evidence for ancient lakes/ponds along the present course of the Colorado River in its Glen Canyon reach and further upstream probably will require considerable care in view of the highly erodible Mesozoic rocks on which these lakes possibly once lay. The gravels in the Wahweap and Bullfrog Basins, presently at or above elevations of

1,250 m, are perhaps the best place to start. A significant challenge will be determining which outcrops represent primary deposition and which have been reworked over the past half million years or more.

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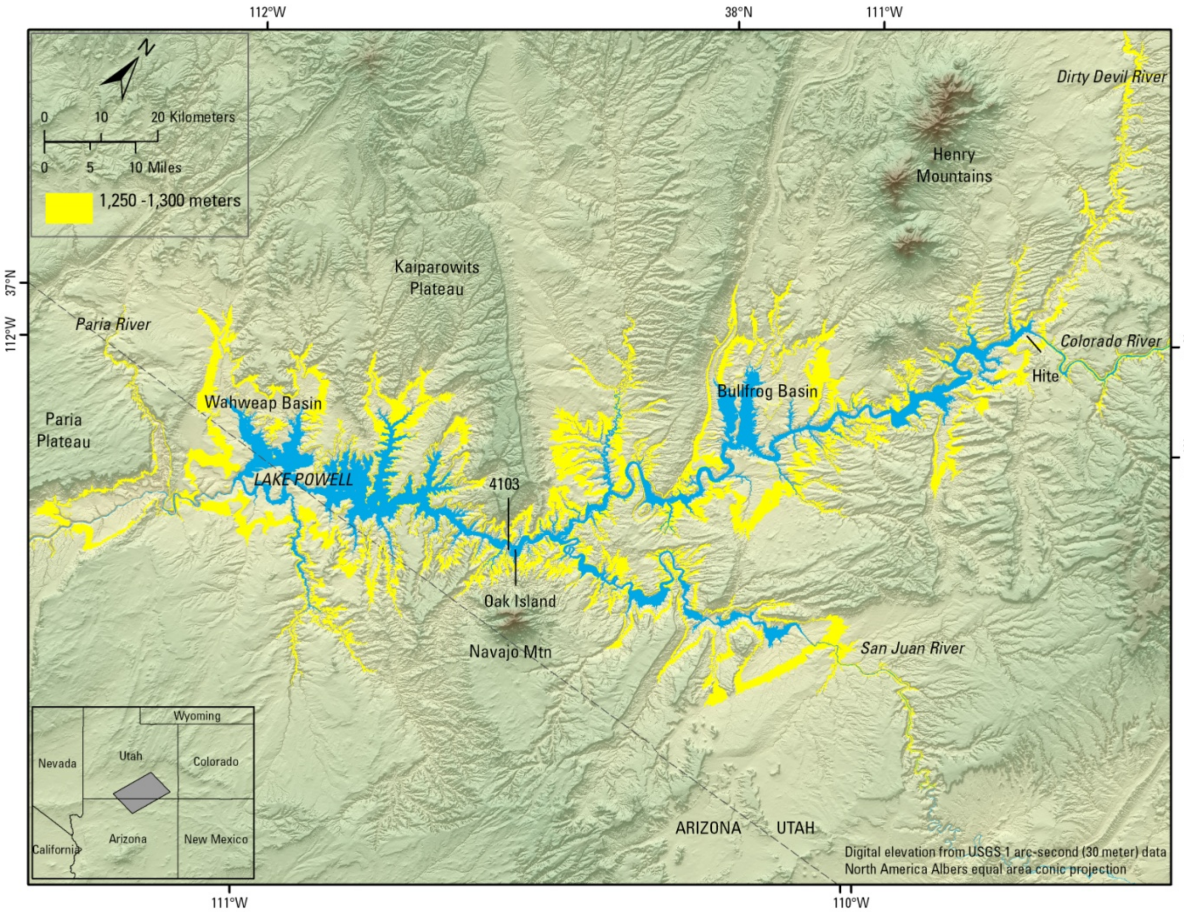

Figure 1. 10-m resolution DEM of the Glen Canyon reach of the Colorado River. Note North arrow; the dashed line from upper left to lower right is the Utah (upper right)–Arizona (lower left) state line. Yellow bands show terrain between the 1,250-m contour, the nominal top of Glen Canyon near Navajo Mountain, and the 1,300-m contour. In the Wahweap (left) and Bullfrog (right of center) Basins, these yellow bands are coincident with outcrops of Canaan Peak gravels derived from the north. Canaan Peak gravels are distinguishable from Colorado River gravels only by the presence of litharenite.

A Working Model for the Evolution of the Grand Canyon/Colorado Plateau Region—Laramide to Present

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In the poster session at the Grand Canyon Workshop Meeting, Flagstaff, Arizona, May 24-26, 2010, we presented eight figures that traced the evolution of Grand Canyon for the time periods of 60 Ma, 50 Ma, 40 Ma, 30 Ma, 17-11 Ma, 11-6 Ma, 6-5 Ma, and 5-0 Ma (fig. 1A-H).We based the paleotopography of our eight figures on: (1) the work of many others, and (2) our own work – specifically on our gravel studies north of the canyon and on our decade-long study of Grand Canyon karst (Hill et. al., 2008; Hill and Ranney, 2008; Polyak and others, 2008; Hill and Polyak, 2010). With the data available for a specific time period, we tried to piece together a likely paleotopographic scenario for that period. Much of this piecing together is speculative, of course, and is akin to Hunt's (1956) model of the early- to middle-Tertiary river and lake system on the southwestern Colorado Plateau. Ours is an attempt to continue Hunt's vision by offering an updated model of evolution for the Colorado Plateau/Grand Canyon region based on new information. We present this effort as a *working model* that can be built upon in the future.

Paleocene Paleotopography (~60 Ma), Figure 1A

For the Late Paleocene to Early Eocene Davis and others (2010) presented evidence for a large river system headed in the arc of the Mojave region that flowed northeast ~700 km into the Uinta Basin of Utah (bold dashed line, fig. 1A) Drainage off the Mogollon Highlands and drainage of an ancestral Little Colorado River east of the Kaibab arch could have flowed into this "California River" system, while drainage west of this system could have flowed along a separate path into Lake Claron west of the Kaibab arch. This western canyon was part of the Laramide "proto-Grand Canyon" proposed by Hill and Ranney (2008), and may represent a very early central Grand Canyon route that the Colorado River occupies today. The locally-derived Robbers Roost Conglomerate (~70-50 Ma) and younger gravels filled paleovalleys on the Coconino Plateau and on the Hualapai Plateau east of the Hurricane fault (Young, 1999).

Early Eocene Paleotopography (~50 Ma), Figure 1B

As the Rocky Mountains rose and adjacent basins developed, large inland lakes and sediment lowlands formed on the Colorado Plateau. The Green River Lake System covered parts of three states and reached its maximum size at ~50 Ma (Smith and others, 2008). Lake Claron could have occupied the southwestern part of this lake system (Hintze, 1988). As in the Paleocene, drainage from the Kingman arch continued to flow northward down a topographic slope and into Lake Claron along the west side of the Kaibab arch, and drainage of the California River continued to flow into the Uinta Basin along the east side of the Kaibab arch. According to Flowers and others (2008), the Upper Granite Gorge area had incised down to Triassic-Jurassic level by ~50-30 Ma, but this "proto-Grand Canyon" existed only in the area of the eastern Grand

Canyon. Coevally and independently, Hill and Ranney (2008) proposed a more extensive "proto-Grand Canyon" that occupied what is now mostly the central and eastern Grand Canyon, from about Mile 80 to Peach Springs Canyon.

In the Eocene the inferred position of the Shivwits scarp was just north of the Hindu Canyon gravel channel (M-H; fig. 1B), and the base of Milkweed Canyon was incised down into the Tapeats Sandstone (Young, 1985). Streams flowing from the Mogollon Highlands and Kingman arch brought arkosic gravels (RG) onto the Hualapai and Coconino Plateaus; i.e., the Music Mountain Formation (~55-45 Ma), which can be traced all the way from the Grand Wash Cliffs to the vicinity of Long Point, with different source areas supplying gravels from east to west.

Late Eocene Paleotopography (~40 Ma), Figure 1C

In the Late Eocene drainage off the Mogollon Highlands and Kingman arch continued to flow north into Lake Claron along the west side of the Kaibab arch. The Green River Lake system was drying up, and the California River of Davis and others (2010) may have become defunct by this time. Hunt (1956) showed a "residual lake" in southeastern Utah in Late Eocene time that was fed by a northward-flowing drainage that may have been the ancestral Little Colorado River. Hill and Ranney (2008) proposed that a "proto-Kanab Creek" flowed north high and along the west side of the Kaibab arch and into Lake Claron at this time. By the Late Eocene, Upper Granite Gorge and a proto-Grand Canyon would have incised even deeper than it had in the Early Eocene.

Rivers incised headward toward the Mogollon Highlands and this drainage became integrated with the Peach Spring drainage system across the old California River route. The main drainage channel incising the Hualapai Plateau was along the northeast-trending Hurricane Monocline, through Truxton Valley, and north along what is Peach Springs Canyon today (Young and Brennan, 1974). The Chino and Aubrey Valleys also supplied gravels onto the Hualapai Plateau from the south. The West Water Limestone was deposited in topographic lows along Milkweed and West Water Canyons. Lake Long Point may (or may not) have been present on the Coconino Plateau. In the Eocene a paleoerosion surface formed on the Hualapai Plateau (Young, 1999). This erosion surface may (or may not) have been contemporaneous with the low-relief Tsaile erosion surface identified by Cather and others (2008) along the flanks of the Chuska Mountains east of the Grand Canyon area.

Oligocene Paleotopography (~30 Ma), Figure 1D

The Oligocene was generally a time of non-deposition in the Grand Canyon region and therefore reconstructions are hard to make. The main sedimentary feature on the Colorado Plateau was the Chuska erg that occupied much of northeast Arizona and northwest New Mexico, and which overlies the Tsaile physiographic erosion surface (Cather and others, 2008). The Buck and Doe Conglomerate formed at this time as a deposit that represented widespread aggradation on the Late Eocene erosion surface across the Hualapai Plateau (Young, 1999).

The Oligocene was also a time of volcanism on and surrounding the northern Colorado Plateau. The Navajo Mountain laccolith probably dates to about 32-23 Ma (Nelson and others, 1992). Uplift of the southern Utah volcanic area caused a reversal of drainage from north into the Lake Claron area to south from Lake Claron toward the Grand Canyon area. Lake Claron ceased to exist except in the Cedar Breaks area, where it remained as a remnant in Oligocene time (Taylor, 1993). Volcanism in southern Utah could have also caused drainage on the east side of the Kaibab arch to shift its position eastward, with an ancestral Little Colorado River feeding an Oligocene lake residual from the Late Eocene lake hypothesized by Hunt (1956). The ancestral Colorado and San Juan Rivers could have joined with the ancestral Little Colorado and could have flowed into this lake during, or somewhat after, this time. The "Crooked Creek river" of Lucchitta and others (2010) may have existed south of the San Juan River, but the extent and source of this river is questionable (G. Billingsley, pers. comm., 2010).

Early reversed drainage southward from the Bryce Canyon area was along the west side of the Kaibab arch, as is documented by Canaan Peak-type gravels at ~7500 ft (Scarborough and others, 2007). This ~7500 ft level may have been part of the ~7500-7600 ft Valencia physiographic surface of Cooley and others (1969). Where this southward drainage flowed to is not known, but Hill and Ranney (2008) speculated that it may have been into Lake Long Point – although no Canaan Peak-type gravels have been found south of the Grand Canyon in this area and Young and Hartman (this volume) consider the Long Point deposits to be Early Eocene in age. The Kingman arch/uplift was probably still in existence to the west and southwest of the Grand Canyon area at least until 30 Ma and perhaps until 17 Ma. These highlands would have directed flow to the east or northeast along the paleochannels of Young and Brennan (1974) and Beard and others (2010) and would have prevented drainage of the Colorado Plateau from flowing west during this time – unless water was already flowing to the south or southwest from the Mogollon slope along the Shivwits scarp area as proposed by Blakey and others (2010).

Middle Miocene Paleotopography (17-11 Ma), Figure 1E

17-16 Ma marks the time of major Basin and Range west-down faulting along the Grand Wash Fault (Faulds and others, 2001). With faulting, water began to flow west towards the Basin and Range. Headward erosion eastward into the Plateau began along the strike valley of the Shivwits scarp, which separated the Hualapai and Shivwits Plateaus (Young, 2001). The "Muddy Creek Formation" along the Grand Canyon mouth area – also called "Rocks of the Grand Wash Trough" by Billingsley and others (2004) – was deposited after Grand Wash faulting began. The ~16-13 Ma fanglomerates were transported from the west or south (or both) and into the trough. As this scarp canyon incised ever eastward, more water accumulated at the mouth of the canyon, and the ~13-11 Ma siltstone-sandstone-gypsum facies of the Muddy Creek Formation was deposited in a series of separate depocenters. The source rocks of this material were the clastics/evaporites of the Kaibab to Supai Formations along the Shivwits scarp (Faulds and others, 1997). The Red Lake halite probably also precipitated at ~13-11 Ma in a flanking basin south of the main evaporative lake where gypsum was precipitating (Faulds and others, 2001); overflow of saline water caused the precipitation of halite. Headward erosion eastward along the Shivwits-scarp canyon had not yet reached the Laramide proto-Grand Canyon channel of Hill and Ranney (2008), so a stream flowing westward from the Kaibab arch could not yet make it all the way to the Basin and Range. This stream water may have ponded in a lake north of Peach Springs Canyon ("lake?", fig. IE).

At 16 Ma deposits of the lower Bidahochi Formation began to accumulate in a local basin occupied by "Lake" Bidahochi. According to Dallegge and others (2001), "Lake" Bidahochi was probably never one large continuous lake but instead consisted of a series of ephemeral lakes. At

16 Ma the ancestral Little Colorado River may have flowed north and the ancestral Colorado River south into the Miocene "Glen Lake" of Hill and others (2008). The ancestral Colorado and Gunnison Rivers flowed west from the Rockies by at least ~11 Ma (Aslan, 2010). The ancestral Green River could have connected with the Snake River in Wyoming, thus accounting for fossil fish of Snake River affinity found in the Bidahochi Formation (Spencer and others, 2008); i.e., the fish could have swam up the Snake to the Green before ~16 Ma when these two rivers became disconnected, then down the Green to the ancestral Colorado River and into Glen Lake, then up the ancestral Little Colorado River and into "Lake" Bidahochi.

Late Miocene Paleotopography (11-6 Ma), Figure 1F

11 Ma was the time when the 16-6 Ma "western Grand Canyon" of Young (2008) and Polyak and others (2008) became integrated from the west side of the Kaibab arch all the way to the Grand Wash Cliffs. This headward eroding western Grand Canyon followed the route of the Laramide proto-Grand Canyon and was occupied by a small river or stream. Headward erosion along a pre-existing proto-Grand Canyon helps solve the "Headward Erosion Problem" posed by Spencer and Pearthree (2001): How could headward erosion have incised eastward to the Kaibab arch and then across it in only 6 Ma? Around 11-12 Ma headward erosion eastward along the Shivwits scarp connected with the Laramide proto-Grand Canyon of Hill and Ranney (2008), thus allowing water to flow all the way from the west side of the Kaibab arch and into the Basin and Range area. Carbonate-rich, high $^{87}Sr/^{86}Sr$, karst spring water – derived from the breaching and dewatering of the Redwall-Muav aquifer by headward erosion as it progressed eastward – flowed as a surface stream to the Grand Wash Cliffs, there to precipitate as the high $^{87}Sr/^{86}Sr$ Hualapai Limestone. Karst water solves the "Hualapai Limestone Problem" posed by Hunt (1974): What was the source of so much carbonate-rich water for the Hualapai Limestone? The reason why the Hualapai Limestone has no clastic delta and is clastic-poor is because the Toroweap, Coconino, and Supai Formations along the proto-Grand Canyon route had *already* been incised during Laramide time; this model thus helps solve the "Muddy Creek Problem" that has troubled geologists for decades. The 16-6 Ma western Grand Canyon continued to erode headward into the west side of the Kaibab arch until it reached Redwall Limestone level along the synclinal axis of the Grandview Monocline. When this happened, a karst-water connection of the eastern and western sections of Grand Canyon occurred *under* the Kaibab arch at ~6 Ma (Hill and others, 2008).

The ancestral Gunnison River connected with an ancestral Colorado River south and west of its present Grand Junction confluence around 11 Ma (Aslan, 2010). "Lake" Bidahochi increasingly broke up into separate ephemeral lakes and playas. The fossil fish of Snake River affinity are found in rocks of ~7 Ma (Spencer and others, 2008). By ~6 Ma the fluvial-eolian upper member of the Bidahochi Formation had been incised by the modern Little Colorado River (Holm, 2001) – perhaps due to the karst connection of the eastern and western sections of Grand Canyon at this time.

Late Miocene-Early Pliocene Paleotopography (~6-5 Ma), Figure 1G

After a karst connection at ~6 Ma, headward erosion proceeded up Little Colorado River Canyon and into the Bidahochi Formation. Headward incision also proceeded up Marble Canyon until a "final connection" was made with Glen Lake at ~5.5 Ma in the area between Navajo Mountain and Fifty-Mile Mountain. This final connection caused the draining of Glen Lake, a reversal of drainage in Marble Canyon, and the integration of the Colorado River from Colorado to the Gulf of California. With the incision of a <6 Ma Grand Canyon, the karst connection section under the Kaibab arch collapsed, deepened, and widened into the Grand Canyon of that section of the Colorado River.

The tectonic opening of the Gulf of California happened at ~7-6 Ma (Dorsey and others, 2007). The lacustrine Bouse Formation was deposited 5.5-5.3 Ma in two paleo-lakes: Lake Mohave and Lake Blythe, which were impounded by paleodams (Spencer and Pearthree, 2005). The southernmost part of the Bouse Formation in the Blythe basin may represent an estuarine environment. Lake Mohave has normal $^{87}Sr/^{86}Sr$ values because this lake was being supplied by Colorado River water *after* integration, in contrast to the Hualapai Limestone whose source was predominantly high $^{87}Sr/^{86}Sr$ karst water derived from the breaching of the Redwall-Muav aquifer *before* the integration of the Colorado River from Colorado to the Gulf of California.

Pliocene Paleotopography to Present Topography (~5-0 Ma), Figure 1H

The integration of the Colorado River through Grand Canyon set off an intense erosion cycle, not only within the canyon itself, but also in the Upper Colorado River Corridor. Over the last 6-5 My, headward incision (knickpoint propagation) has proceeded up the Little Colorado River, the Colorado River in Glen Canyon (and above Glen Canyon), the San Juan River, and the Green River, thus creating the deep, narrow, "young" canyons that are present today along these rivers and tributaries.

Large tributaries to the Colorado River within Grand Canyon (Havasu-Cataract, Kanab, Little Colorado River) were already forming before the Colorado River erosion cycle and thus appear older than other tributaries incised during the present Colorado River erosion cycle. Over the last 5 Ma or so the paleolakes along the Lower Colorado River Corridor became drained as the Colorado River established its final course to the Gulf of California.

While most of the erosion of Grand Canyon by the Colorado River has occurred over the last ~5-0 Ma, this does not diminish the role that the two earlier episodes – the "Laramide proto-Grand Canyon" and "16-6 Ma western Grand Canyon" – played in establishing the route that the Colorado River now takes through the canyon. It is proposed that the route of the central and eastern Grand Canyon was established way back in the Laramide; then, starting at about 11 Ma, the headward-eroding Shivwits scarp section became integrated with this combined proto-Grand Canyon/western Grand Canyon; at 6 Ma, the section of Grand Canyon east of the Kaibab arch became connected to the 16-6 Ma western section via a karst connection (Hill and others, 2008); and at ~5.5 Ma the "final connection" was made with Glen Lake, thus allowing Colorado River gravels to travel through Grand Canyon and first appear at the mouth of the canyon/Grand Wash Cliffs at ~5.5 Ma.

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Figure 1 (below). Proposed drainage evolution of the Colorado Plateau with respect to the Grand Canyon area in the time frame of 60 Ma, 50 Ma, 40 Ma, 30 Ma, 17-11 Ma, 11-6 Ma, 6-5 Ma, and 5-0 Ma. F = Flagstaff, P = Prescott, PS = Peach Springs, LV = Las Vegas, K = Kanab, SL = Salt Lake, GJ = Grand Junction, Gn = Gunnison, W = Winslow, Y = Yuma, T = Tucson, LF = Lees Ferry, LP = Long Point, BC = Bryce Canyon, CP = Canaan Peak area, C = Confluence, aCoR = ancestral Colorado River, aSJR = ancestral San Juan River, aLCoR = ancestral Little Colorado River, aGR = ancestral Green River, aGnR = ancestral Gunnison River, CRr = Crooked Ridge river, UGG = Upper Granite Gorge, GWC = Grand Wash Cliffs, SS = Shivwits Scarp, RG = rim gravels, M-H = Milkweed-Hindu Canyon, WK = West Kaibab arch Canaan Peak-type gravels, Lake LP = Lake Long Point, Lake WW = Lake West Water, TV = Truxton Valley, C-AV = Chino-Aubrey Valley, H-C Ck = Havasu-Cataract Creek, JC = Johnson Creek, CK = Cedar Knoll, LCK = Little Cedar Knoll, GN = Goosenecks, RL halite = Red Lake halite, FC5.5 Ma = Final connection(?) 5.5 Ma, pd = paleodam. Bold dashed line = California River. Between the red lines = the proposed section of canyon that the Colorado River still takes today. The term “ancestral” with regard to rivers means *before* the Colorado River became integrated through Grand Canyon from Colorado to the Gulf of California at ~5.5 Ma. Base maps are from Blakey and Ranney (2008). We thank Ron Blakey for giving us permission to use his paleotopographic maps.

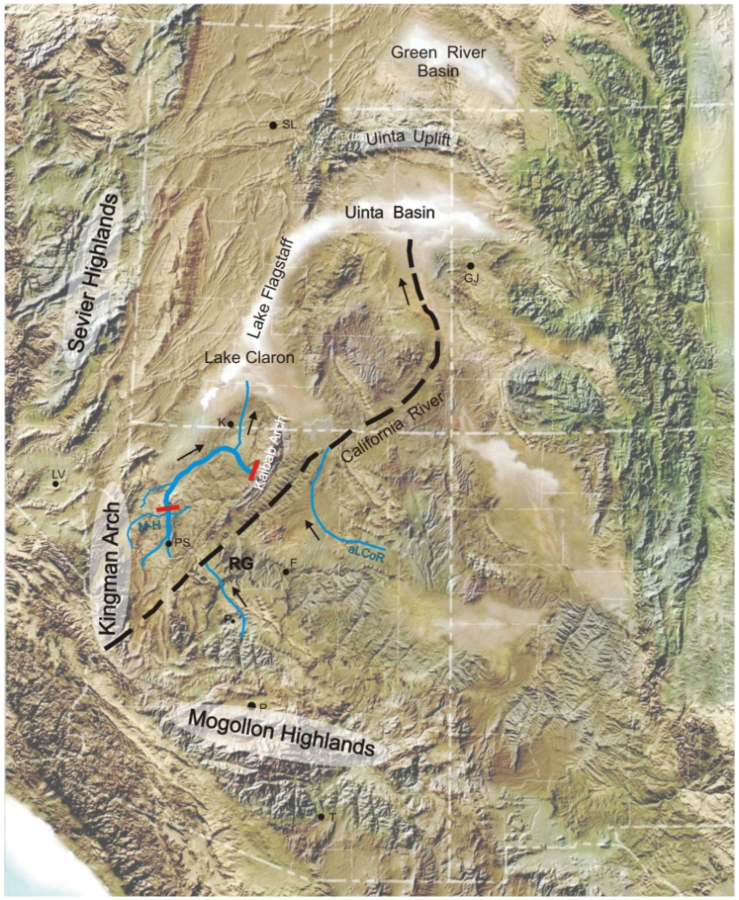

Fig. 1A. Paleocene paleotopography, ~60 Ma.

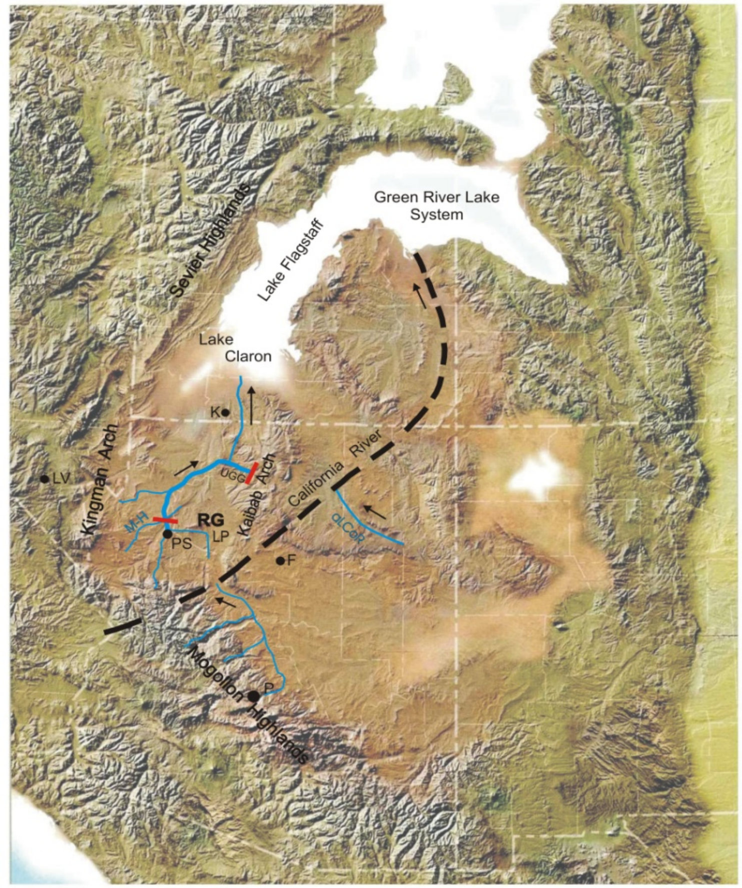

Fig. 1B. Early Eocene paleotopography, ~50 Ma.

Fig. 1C. Late Eocene paleotopography, ~40 Ma.

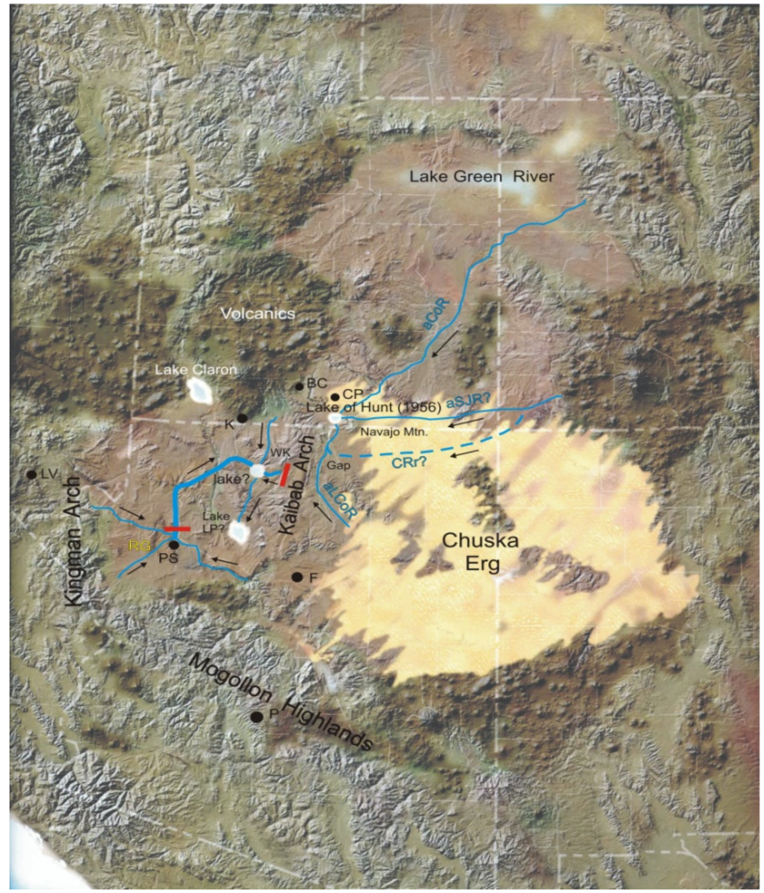

Fig. 1D. Oligocene paleotopography, ~30 Ma.

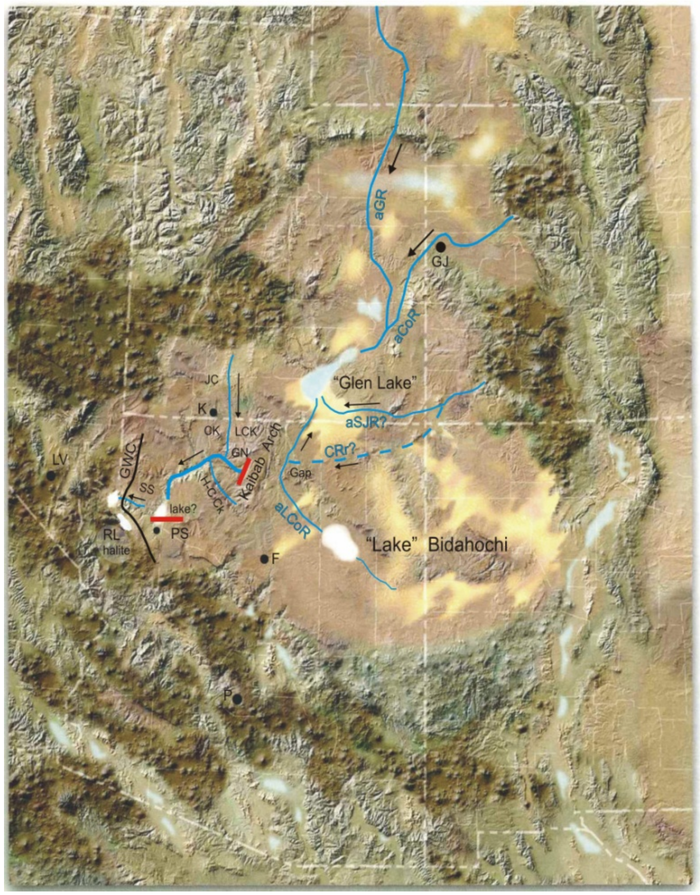

Fig. 1E. Middle Miocene paleotopography, 17-11 Ma.

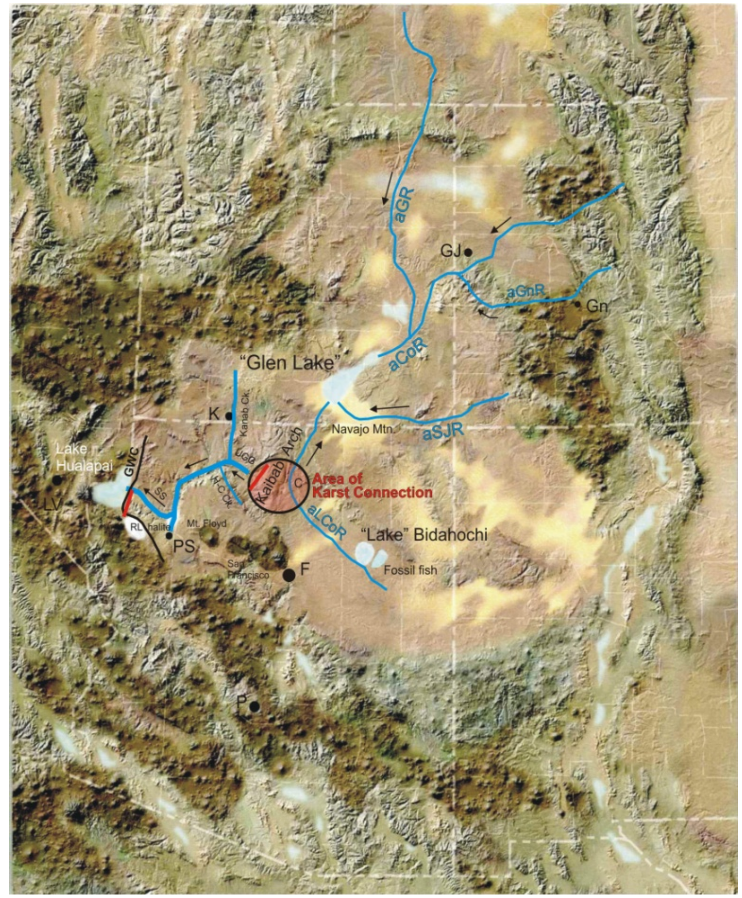

Fig. 1F. Late Miocene paleotopography, 11-6 Ma.

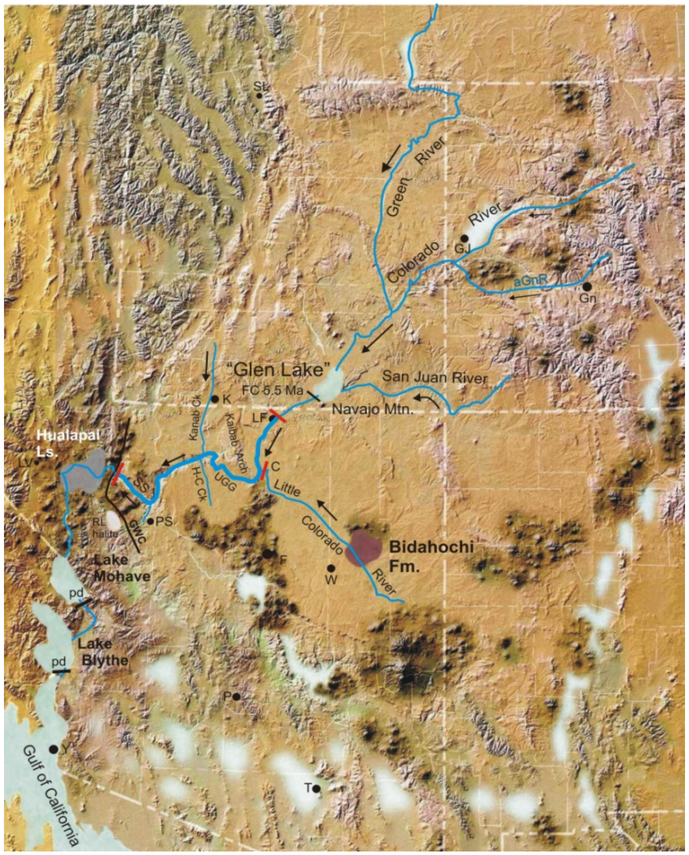

Fig. 1G. L.Miocene-E.Pliocene paleotopography, ~6-5 Ma.

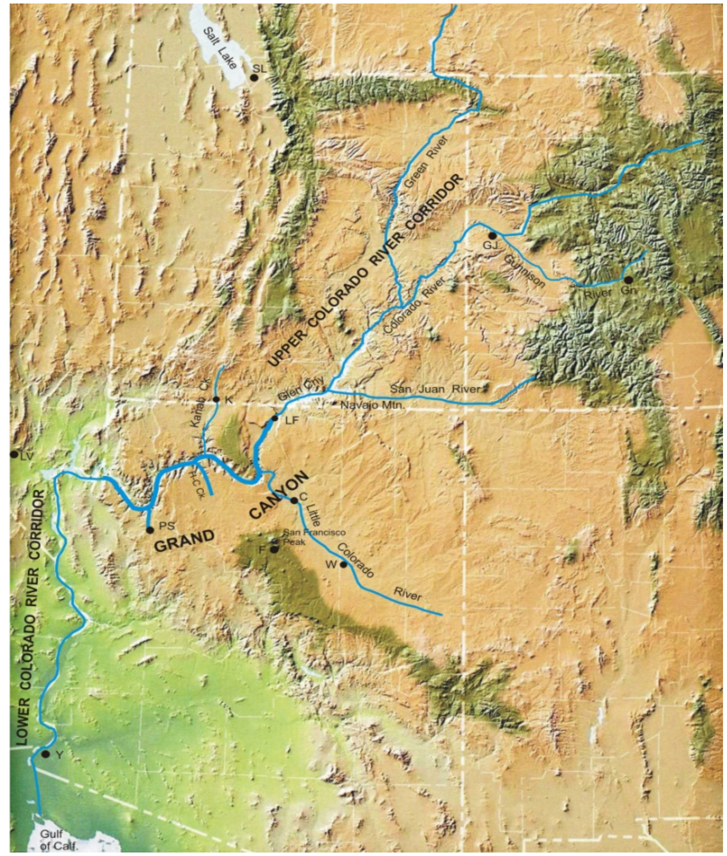

Fig. 1H. Pliocene paleotopography to Present, ~5-0 Ma.

Mio-Pliocene Erosional Exhumation of the Central Colorado Plateau, Eastern Utah—New Insights from Apatite (U-Th)/He Thermochronometry

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The landscape evolution of the Colorado Plateau has been studied by scientists for over a century, yet its late Cenozoic erosional and geomorphic history remains poorly understood. This study uses apatite (U-Th)/He thermochronometry to investigate Neogene erosional exhumation in the center of the Colorado Plateau in eastern Utah (fig. 1). Apatite (U-Th)/He thermochronometry (AHe) has been shown to be a powerful tool to reconstruct long-term thermal histories (e.g., House and others, 1997; Wolf and others, 1997; Farley, 2000; Stockli and others, 2000; Reiners and others, 2000; Farley and Stockli, 2002; Ehlers and Farley, 2003) and its low-temperature sensitivity is ideally suited to constrain the magnitude, timing, and spatial patterns of erosion in this area. This study presents new thermochronometric data (93 samples, 517 aliquot ages) from a combination of cores and surface samples from the Monument Uplift, Canyonlands, Book Cliffs, and Uinta Basin regions. The combination of AHe thermochronometry and thermal modeling (Hager and Stockli, 2009) enables us to elucidate the thermal and erosional history of the central Colorado Plateau and to reconstruct the long-term landscape evolution of this region (<10 Ma).

In the Monument Uplift region, AHe results from core samples and surface samples exhibit two different trends. Surface sample ages from the Monument Uplift exhibit much scatter over similar elevations between 13–50 Ma (Stockli and others, 2002), yet core ages change little with depth or elevation 3–12 Ma. The surface sample ages reveal slower cooling in the HePRZ for at least 40 M.y., between 10–50 Ma. Similar core ages at varying elevations indicate significant late Miocene to Pliocene cooling and exhumation. The thermal modeling for Monument Uplift show maximum reheating temperatures of the uppermost modeled sample that range between 55–70°C and the onset of rapid cooling between 4–10 Ma. Based on a geothermal gradient of 30°C/km (Blackett, 2004) and mean annual surface temperature of 10°C, these HePRZ temperatures correspond to 1.5–2.0 km of overburden removal in the Monument Uplift region since late Miocene time.

In the Canyonlands region, 13 surface samples were collected from a vertical transect along the Shafter Trail from the top of the Island in the Sky District to the Colorado River level, spanning about 700 m. The samples yielded a broad spread of Neogene to late Paleogene ages between 4.5 ± 0.4 and 41.5 ± 3.3 Ma (n=61), with an overall trend of increasing ages with increasing elevation. The Canyonlands region is protected by the National Park Service and no wells are located within the park, however two boreholes (depths ~800-950 m) were sampled immediately north of Canyonlands National Park. The three core samples analyzed from the Canyonlands region generated mid to late Miocene AHe ages ranging from 4.4 ± 0.3 to 11.0 ± 0.2 Ma (n=11).

Thermal modeling results from the Canyonlands area show maximum reheating temperatures mainly between 50−70°C and show the onset of cooling at 4−5 Ma. Using a geothermal gradient of 20°C/km (Blackett, 2004) and this range of maximum temperatures, 2−3 km of erosion is calculated for the Canyonlands region since the late Miocene.

Across the Book Cliffs region, a total of 34 samples were gathered along three transects: (1) a transect across the Book Cliffs and Roan Cliffs in Hay Canyon near the Utah-Colorado border, (2) a transect of the central Book Cliffs in Sego Canyon, and (3) a vertical transect in the Blaze Canyon area of the central Book Cliffs. At Hay Canyon, 12 samples were collected from the top of the Roan Cliffs to the base of the Book Cliffs escarpment, with about 900 m of vertical elevation spread over 21 km horizontal distance. The surface samples from this transect have a wide range of Eocene to early Pliocene ages. AHe ages ranged from 5.5 ± 0.6 − 29.1 ± 4.3 Ma (n=63). In Sego Canyon, the 13 surface samples yielded AHe ages ranging from 8.1 ± 0.5 − 51.8 ± 3.1 Ma (n=59). In the Blaze Canyon region seven surface samples were collected up the front of the Book Cliffs escarpment, totaling 400 m of relief. These samples yielded AHe ages ranging from 2.1 ± 0.1 − 35.7 ± 2.1 Ma (n=29). A majority of the cores collected are from the Book Cliffs, especially the eastern Book Cliffs near Hay Canyon, an intensely explored region with many old oil exploration wells. Twenty-two shallow core samples (<300 m depth) generated a broad span of AHe ages 1.2 ± 0.1 − 55.8 ± 5.2 Ma (n=111). Deeper core samples (depths 900-1400 m) were less available due to an abundance of Paradox halite and carbonate, but three samples produced AHe ages ranging from 2.4 ± 0.3 − 7.4 ± 0.2 Ma (n=8). Thermal modeling shows maximum burial temperatures for surface samples range from 40−60°C in Hay Canyon with erosional exhumation starting 5−10 Ma. Using the average gradient of 26°C/km, 1.2−1.9 km of sediment has been removed since the late Miocene. In Sego Canyon, thermal modeling histories show maximum burial temperatures between 45−65°C and erosional exhumation starting between 5−8 Ma. Using a geothermal gradient of 32°C/km (Blackett, 2004), 1.1−1.7 km of erosion is determined in the Sego Canyon region. In the Blaze Canyon region, maximum reheating temperatures range from 50−70°C prior to rapid erosional exhumation at 5 Ma. Assuming a geothermal gradient of 34°C/km (Blackett, 2004), 1.2−1.8 km of erosion has occurred in the central Book Cliffs.

Wells from the Uinta Basin targeted early Cretaceous strata and thus these wells commonly were very deep (~1300−2700 m). Approximately half of the aliquots from Uinta Basin samples yielded ages between 1.0 ± 0.1 − 4.2 ± 0.6 Ma (n=21). The shallowest core sample (~250 m elevation) returned ages 3.4 ± 0.2 − 6.9 ± 0.4 Ma (n=3). The remaining aliquots (n=22) had very low amounts of He (<0.1 nmol/μg) and AHe ages <1 Ma. Based on modern geothermal gradients of 30 °C/km in the Uinta Basin and a mean surface temperature of 10 °C, current temperatures at the depth of most of these samples (>2000 m) range from 70−90 °C. At these temperatures, AHe ages should be completely reset and AHe ages <1 Ma are consistent with this assessment. This indicates the depth to the base of the modern HePRZ in the Uinta Basin is approximately 2300 m.

In summary, all apatite (U-Th)/He (AHe) ages from surface and core samples are younger than stratigraphic ages, suggesting complete or partial thermal resetting after deposition and burial. Core samples (depths >1 km) have proven critical to this study and record the onset of significant

Mio-Pliocene cooling and exhumation at ca. 6 Ma. Shallower cores and surface samples have a broad spread of Eocene to late Miocene ages (5−55 Ma), and indicate residence in the helium partial retention zone (HePRZ).

AHe data combined with thermal modeling provide powerful constraints for erosion of the central Colorado Plateau and shows that its erosional history varies from that of the surrounding edges of the plateau. Significant rapid erosion has dominated in the eastern Utah area of the central plateau since ca. 6 Ma. Since the latest Miocene to early Pliocene, 1−3 km of sediment overburden has been removed from the central plateau, whereas relatively little erosion has occurred along the periphery, confirming a predicted bull's eye trend in the spatial distribution of erosion (Pederson and others, 2007). The spatial distribution of erosion appears to have a concentric pattern, with the greatest amount of erosion in the Canyonlands region, slightly less in Monument Uplift and Book Cliffs region, and far less in the Uinta Basin. Overall these estimates further constrain previous erosion estimates for the central Colorado Plateau and are consistent with many other estimates for the area (e.g., Dumitru and others, 1994; Nuccio and Condon, 1996; Stockli and others, 2002; Pederson and others, 2002b; 2007), but also add important temporal constraints to the landscape evolution of this region.

We argue that erosion in the central Colorado Plateau is not a simple consequence of surface uplift and is temporally decoupled from numerous proposed regional uplift mechanisms (e.g., crustal thickening (McQuarrie and Chase, 2000), delamination of mantle lithosphere (Spencer, 1996), anomalous temperature or chemistry of mantle lithosphere (Humphreys and others, 2003; Roy and others, 2009), and buoyant asthenospheric upwelling (Moucha and others, 2009; Liu and Gurnis, 2010)). Conversely, we attribute Mio-Pliocene erosion to a synthesis of driving forces at 6 Ma: drainage integration of the Colorado River off the plateau, the intensification of the southwest monsoon climate, and the opening of the Gulf of California. These synchronous events combined to produce up to 3 km of erosion in the central plateau and resulted in the evolution of the remarkably young, central Colorado Plateau landscape.

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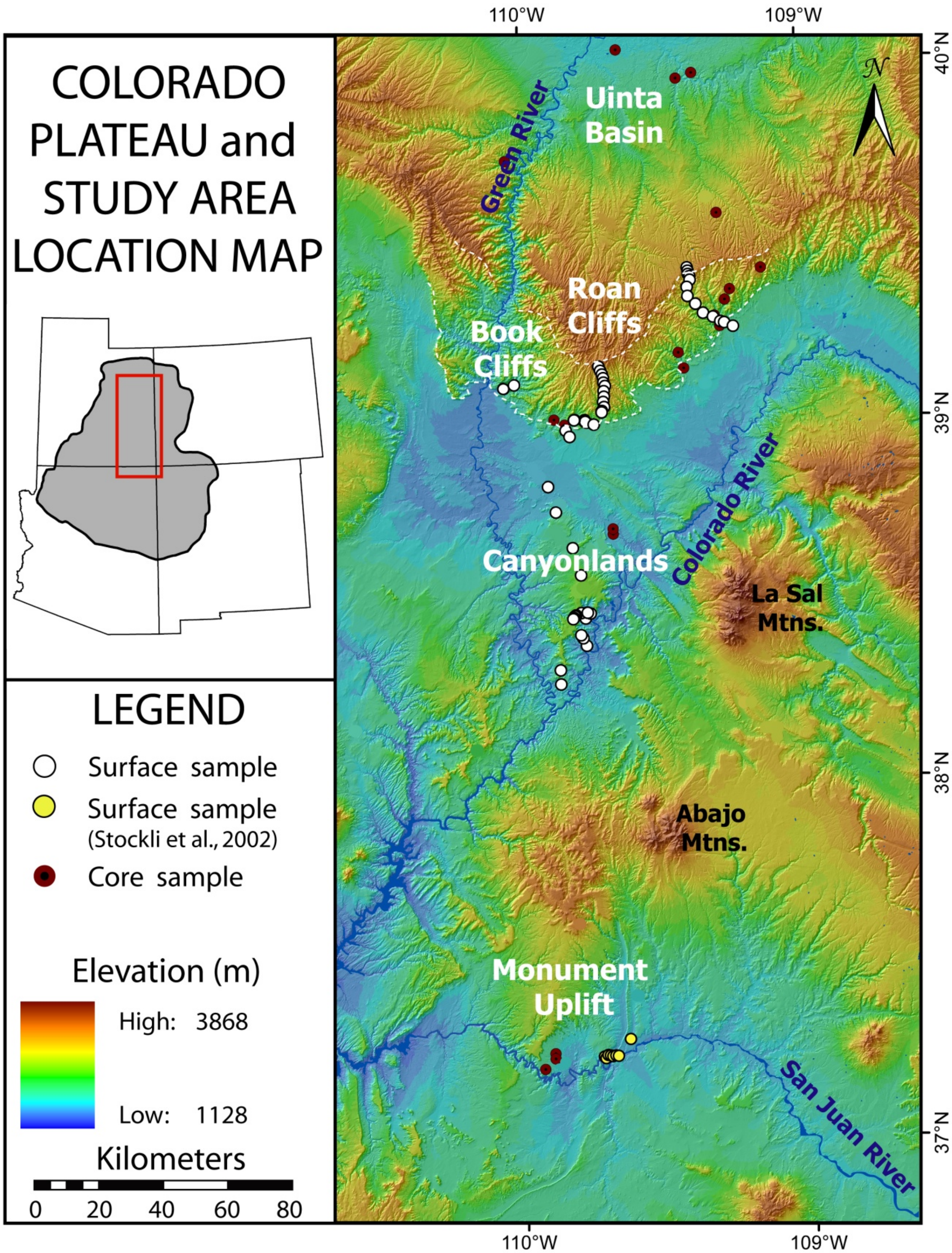

Figure 1. Figure inset shows the Colorado Plateau geographic province in the western United States with location the study area in eastern Utah (red box). Digital elevation model (DEM) of eastern Utah shows sample locations from the four main focus areas: Monument Uplift, Canyonlands, Book Cliffs, and Uinta Basin.

Robust Geologic Evidence for latest Miocene-earliest Pliocene River Integration via Lake-spillover along the lower Colorado River—Review and New Data

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Over the decade since the last Grand Canyon symposium in 2000 our geologic mapping efforts along the lower Colorado River have uncovered a trove of evidence that is best explained by the lake-spillover model of river integration (fig. 1). In particular, new sedimentologic, stratigraphic, geochronologic, and geomorphic evidence in Cottonwood Valley (CV) and Mohave Valley (MV) near the junction of California, Nevada, and Arizona are consistent with this sequence of events (oldest to youngest; key geologic units or features indicated in italics; all elevations in meters above sea level):

1. Late Miocene sedimentation in two basins separated by a bedrock divide, with the floor of CV (>200 m) substantially higher than MV (<140 m): *Lost Cabin beds in Cottonwood Valley; Fanglomerate and Newberry gravel in Mohave Valley.*

2. Expansion of fine-grained axial valley deposition in CV in latest Miocene time: *Lost Cabin beds.*

3. Rapid filling of CV with a moderately deep lake to >360 m asl after 5.6 Ma, presumably with proto-Colorado River water; *Upper Lost Cabin beds, with local marl.*

4. Lake spillover into MV and erosion of a bedrock divide: *Pyramid gravel.*

5. Filling of both valleys with one deep lake to 555 m: *Bouse Formation.*

6. Southward overflow of the deep lake into the Chemehuevi Valley-Lake Havasu area and erosion of the former divide; eventual drainage of the MV-CV lake and erosion of Bouse deposits and older units: *unconformity between Bouse and Bullhead.*

7. Arrival of distinctive Colorado River sand and gravel in both basin axes with voluminous reworked local sediments: *lower Bullhead alluvium.*

8. Massive Colorado River aggradation resulting in valley filling to ~400 m, culminating about 3.5 to 4 Ma: *Bullhead alluvium.*

Late Miocene, predominantly undeformed alluvial fan and basin axis deposits exposed in CV and northern MV (fig. 2) preclude the presence of a through-going river at that time (e.g., Metzger and Loeltz, 1973). The central part of CV contains the Lost Cabin beds (House and others, 2005, 2008), a sequence of gently dipping to flat-lying interbedded sandstone and mudstone that is similar to younger parts of the Muddy Creek Formation and especially the Willow Beach beds (Amoroso and Felger, 2011) in basins to the north. The 5.59 ± 0.05 Ma Wolverine Creek tephra has been identified at two localities in the upper part of the Lost Cabin bed section (House and others, 2008), indicating approximate temporal equivalence with the Hualapai Limestone (Spencer and others, 2001) and possibly with younger parts of the Muddy Creek Formation.

Distinctive gravel deposits derived from granitic rocks of the Newberry Mountains to the west and volcanic rocks of the Black Mountain to the east grade laterally and vertically into the Lost Cabin beds, which filled the valley axis and troughs between large alluvial fans emanating from the bounding ranges. Recently discovered Lost Cabin beds in central CV are lower in altitude than those to the south, consistent with a depocenter in central CV, and progressive valley filling and areal expansion of fine-grained sedimentation into southern CV in latest Miocene time. Similarities between the Lost Cabin beds and the upper Muddy Creek Formation may be simply the result of similar depositional conditions, but gradual expansion of Lost Cabin deposition suggests an incipient hydrologic connection between CV and the basins upstream, such as the Willow Beach area as described in Amoroso and Felger, 2011).

The Lost Cabin beds and temporally equivalent fanglomerates in CV are overlain by the Bouse Formation. In the valley axis, Bouse deposits generally rest disconformably on Lost Cabin beds, but interfinger with them locally. Away from the valley axis, Bouse deposits rest disconformably on paleo-alluvial fan and bedrock surfaces as high as 555 m. Bouse Formation outcrops are generally thin and consist of basal carbonate or tabular sandstone and fine gravel layers, locally overlain by greenish marl. Some of the highest Bouse outcrops consist of cross-bedded coarser sand and fine gravel—likely littoral deposits. All preserved Bouse sections in CV are <10 m thick, and their upper contact is generally an unconformity overlain by locally derived fanglomerate. Across northern Mohave Valley, late Miocene deposits are locally derived fanglomerates derived from three distinct sources: Newberry granite to the east, Black Mountain volcanic rocks to the west, and Proterozoic megacrystic granite to the north. The latter rocks form the bulk of the hills that divide CV and MV. Locally derived late Miocene alluvial fan deposits are also exposed in many localities around the eastern, western and southern margins of Mohave Valley. The alluvial fans and presumably an axial drainage system fed into a depocenter in central or southern MV, where temporally equivalent fine-grained deposits are surely in the subsurface. Thus, CV and MV had separate catchments in late Miocene time.

A key fluvial boulder conglomerate in the axis of northern MV indicates that the linkage between the valleys developed through a process of divide overtopping and catastrophic flooding. The boulder conglomerate is dominated by megacrystic granite derived from the paleodivide between the valleys (the Pyramid gravel of House and others, 2005, 2008), and occupies a deep, wide, south-sloping erosional paleochannel carved in Newberry fanglomerate. Pyramid gravel deposits are conformably overlain by basal carbonate deposits of the Bouse Formation at ~200 m. Thus, quiet-water Bouse deposition immediately followed flooding associated with erosion of the paleodivide. Bouse deposits are exposed in many other places in MV. Most common outcrops consist of the basal carbonate unit less than 2 m thick resting on paleo-alluvial fan surfaces, or locally on bedrock; these are typically disconformably overlain by local fanglomerate. In central and southern MV, some Bouse deposits are much thicker and consist of mud to fine sand above the basal carbonate unit. In southern MV, the basal carbonate rests on alluvial fan deposits to as high as 170 m; the base of Bouse outcrops closest to the valley axis is not exposed but must be less than 140 m asl. The southward decreasing elevation of the base of the Bouse Formation is consistent with a pre-Bouse depocenter at the southern end of Mohave Valley.

The next younger set of deposits in both valleys are distinctive quartz-rich sand, rounded gravel, and mud of the through-flowing Colorado River (the alluvium of Bullhead City or Bullhead alluvium; House and others, 2005, 2008). In the valley axis, Bullhead alluvium filled and buried erosional topography cut into Bouse deposits and older units down to ~150 m asl. Evidently, the CV-MV lake basin was drained and partially eroded prior to arrival of Colorado River bedload, which implies that the southern paleodivide was eroded down to a level close to the modern Colorado River. Higher on the valley margins, Bullhead alluvium interfingered with local fanglomerate as the valleys filled with sediment. The highest Bullhead deposits are about 400 m, and outcrops of the 3.6–4.2 Ma Lower Nomlaki tephra in tributary fan deposits at altitudes from 360–390 m asl are overlain by Bullhead alluvium. The 3.29 ±0.05 Ma Nomlaki tephra (note: distinct from the Lower Nomlaki) crops out at 350 m in younger tributary gravels that overlie an erosional unconformity cut on Bullhead sediments (House and others, 2008). Bullhead aggradation culminated by 3.5–4 Ma. This major phase of early river aggradation may owe to high rates of erosion by the Colorado River and tributaries in the Grand Canyon area.

Is downstream-directed river integration via-lake spillover a plausible mechanism? The Owens (Phillips, 2008), Mojave (Meek, 1989; Miller, 2005), and Amargosa (Menges, 2008) rivers provide examples of downstream-directed integration through once-enclosed basins continued to regions of insurmountable topographic enclosure. The Rio Grande provides an example of complete downstream-directed integration to the sea (Connell and others, 2005). The balance between water and sediment inputs and topographic impediments determines the extent, continuity, and persistence of integration. The lower Colorado River integrated this way because of the juxtaposition of a large highland watershed (Southern Rockies and Colorado Plateau) against the lower, topographically complex Basin and Range. The emergent Gulf of California ultimately provided the impetus to develop a conduit between the upper basin and the sea.

Perhaps the most challenging questions involve the specific hydrologic and geologic circumstances on the Colorado Plateau and areas farther upstream that acted to initiate the process of integration.

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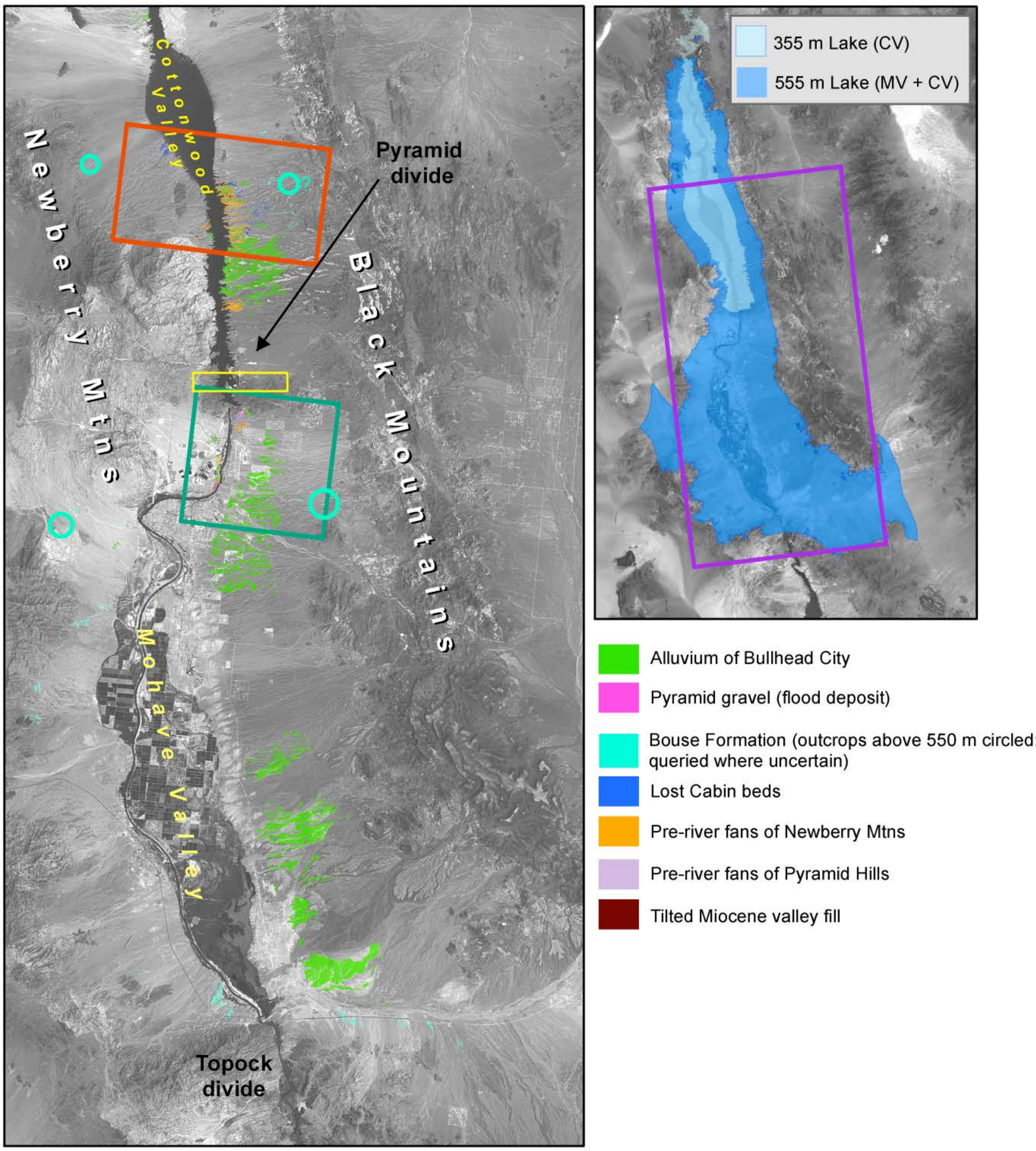

Figure 1a.
Map (left) showing key outcrops in Cottonwood Valley (CV) and Mohave Valley (MV) that document the arrival and early aggradation of the lower Colorado River; and map (right) showing likely extents of the two successive lakes related to the origin of the river.

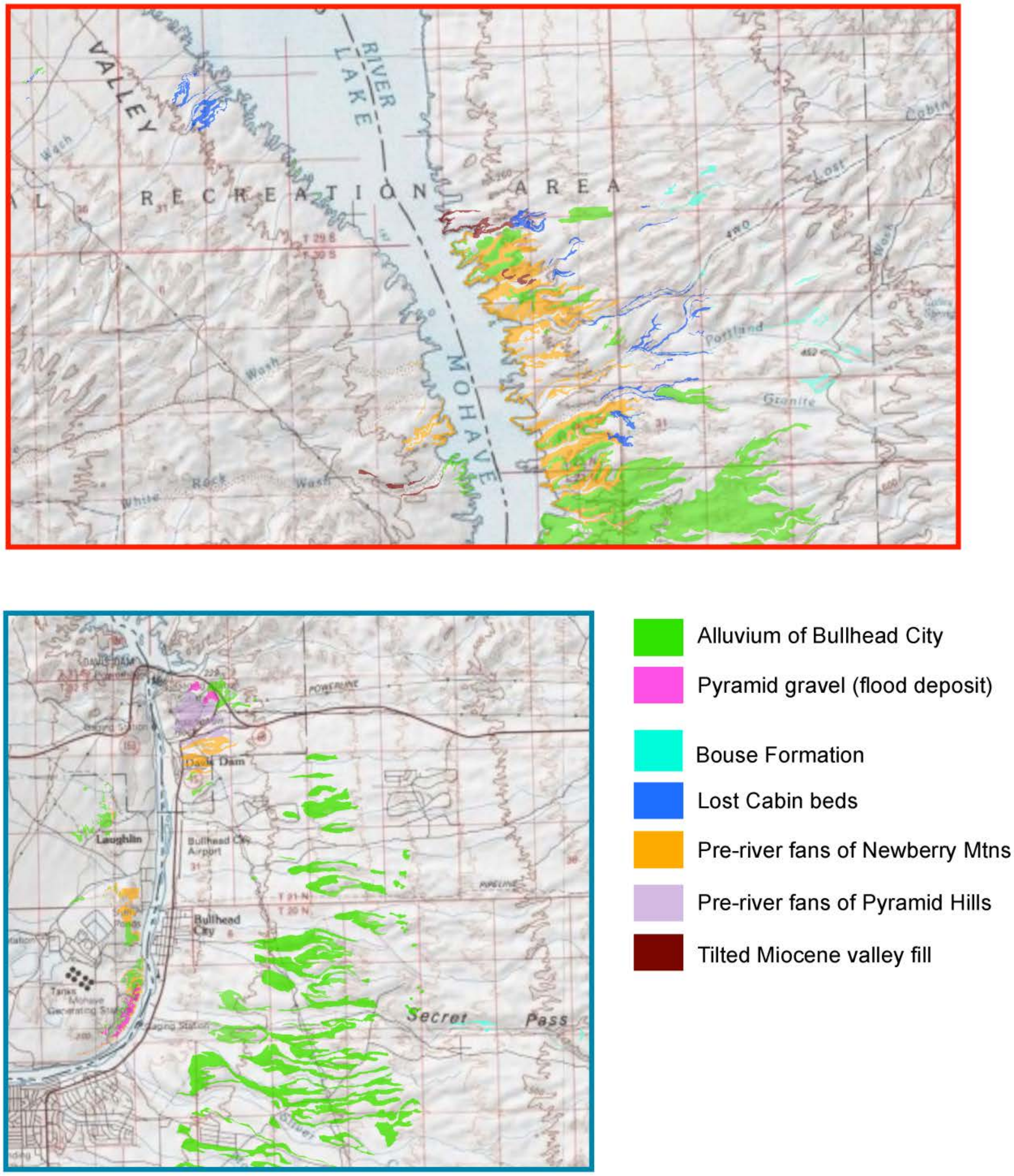

Figure 1b.
Detailed maps of key areas shown in Figure 1a.

Pliocene Aggradational Sequence of the Lower Colorado River in Longitudinal Profile

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A Pliocene aggradation of Colorado River sandstone and conglomerate, exemplified by the pre-3.3 Ma alluvium of Bullhead City of House and others (2005, 2008), followed deposition and subsequent incision of the Bouse Formation. This sequence is exemplified by the pre-3.3 Ma alluvium of Bullhead City of House and others (2005, 2008). Probable correlatives of the Bullhead City unit occur all along the river's course through the Basin and Range province, from near the mouth of Grand Canyon, Arizona, downstream to the U.S.-Mexico border. The stratigraphic sections in individual reaches vary as much as 170–330 m in elevation range, indicative of large original thickness. The voluminous aggradation(s) built up wide braid plains following deep incision by the river into the Hualapai and Bouse formations. A longitudinal profile of the top of this aggradational sequence (excluding folded and faulted sections near Lake Mead and near the San Andreas fault) has an average downstream slope of 0.0003–0.0008 (fig. 1). This is steeper than of the historic river, and may be depositional or have a tectonic component. Whether the sequence (excluding in the delta) generally thickens upstream, which might imply aggradation driven by a high sediment load in relation to discharge, remains to be tested. Downstream of the southernmost canyon through the Chocolate Mountains, the upper part of the deposits facing the open delta near Yuma, Arizona, is high as 230 m asl. High elevation of the upper part of the Pliocene sequence here, concordant in grade to similar sections upstream, could be explained by (1) damming behnd an unknown downstream barrier in the delta area; (2) regional uplift; or (3) grading of the Pliocene sequence toward a distant base level, implying a delta much more extensive than the modern delta. The latter interpretation seems likely. In the delta area southwest aof the San Andreas fault systyem, the sequence is thickest; is deeply buried, indicating subsidence toward the Gulf of California-Salton trough; and contains interfingered fluvial and marine beds (Olmsted and others, 1973), indicating deposition at the delta front. This stratigraphic record suggests that, as the Pliocene aggradation continued, the delta prograded extensively seaward into a tectonically subsiding basin, that of the Gulf of California and Salton Trough).

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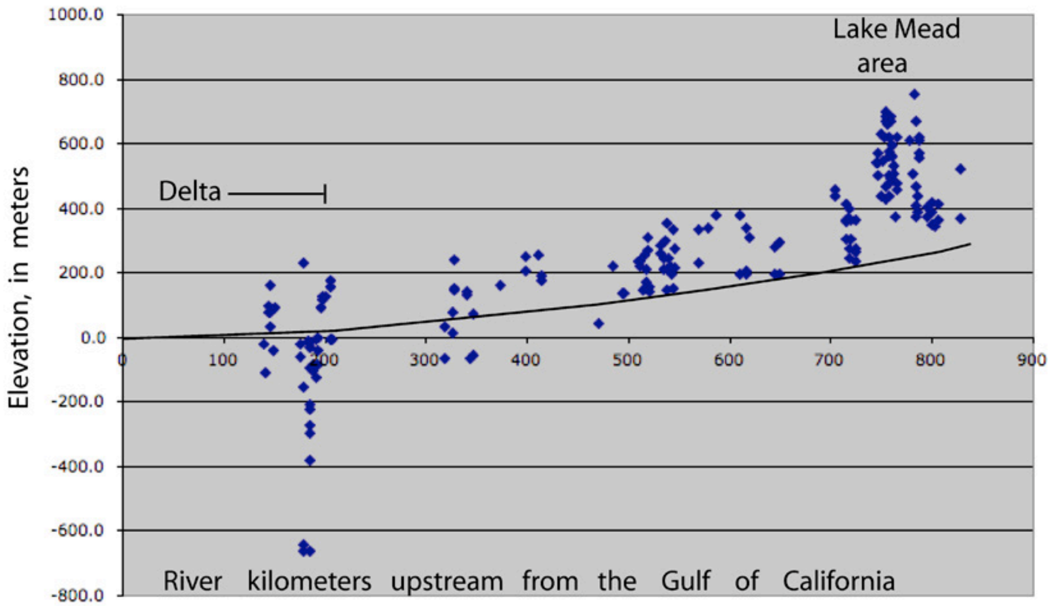

Figure 1. Longitudinal profile of Pliocene Colorado River sediments (points) compared to the historical Colorado River (curve).

Boulders Deposited by Pliocene and Pleistocene Floods on the Lower Colorado River

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Boulder gravels in several parts of the stratigraphic record of the lower Colorado River downstream from the Grand Canyon record significant floods (fig. 1). The boulder gravels vary from Pliocene to late Quaternary in age. They are much coarser than the sandy to gravelly bed of the modern river. Many of the boulder deposits are in confined canyon reaches where high flow velocities can occur during high discharges. Most of the boulders are locally derived, likely from river reworking of tributary debris flows or debris-flow deposits.

A single late Pliocene (or early Pleistocene?) flood may be recorded by boulder conglomerate in Arizona and California over a Colorado River reach 22 km long in lower Mohave Valley, Topock Gorge, and upper Chemehuevi Valley (fig. 2). The conglomerate is deeply inset into, and carries sandstone clasts derived from, the lower Pliocene alluvium of Bullhead City of House and others (2005). The boulder conglomerate records a cycle of degradation-aggradation-degradation between early Pliocene time and the late Pleistocene age of the Chemehuevi Formation of Longwell (1963). The well sorted and imbricated deposit lacks internal layering and fines upward to cobbles. Its total projected ~45 m thickness includes the fill of a central channel at least 20 m deep and lateral deposits that grade gently onto adjacent paleovalley slopes as far as 1–2 km from the central paleochannel. The channel thalweg slopes southward roughly parallel to the historical river gradient but is 20–30 m higher. Coarse, rounded cobbles of far-traveled quartzite are common, but larger clasts ($\geq$1 m) are locally derived volcanic rocks, gneiss, and granite. The coarsest clasts are at the head of Chemehuevi Valley, where rounded and river-sculpted granite clasts as wide as 3 m, derived from the nearby Chemehuevi Peak Granodiorite, likely were reworked from fanglomerate substrate, or from a debris flow into the river. The deposit fines downstream to coarse cobbles over a distance of 5 km. If the deposit correlates with the bouldery conglomerate of Laughlin of House and others (2005) 40 km upstream, it could record a regional flood, rather than the local breakout flood that House and others suggested for the Laughlin unit.

Pliocene fluvial conglomerate in the Lake Mead area contains far-traveled quartzite cobbles and larger, locally derived granite and gneiss boulders, some in Virgin Canyon as large as 5 m (Howard and others, 2003). A layer of blocks derived from Hualapai Limestone is interbedded in the middle of a fluvial pebble-cobble gravel section west of Temple Bar and likely records a tributary debris flow into the river.

Extrapolation of drilling logs for the Parker Dam site in Aubrey Canyon (Berkey, 1935) suggest two buried boulder conglomerates, one in the lower part of the 80-m-thick canyon fill (Pliocene?) and a higher one that may be Quaternary (fig. 3). Newberry (1861) found a mammoth tooth in boulder conglomerate underlying fine-grained beds of the upper Pleistocene Chemehuevi Formation of Longwell (1963) in Cottonwood Valley. Late Pleistocene, post-Chemehuevi Formation terrace gravels in Topock Gorge contain local and far-traveled boulders derived from quartzite, porphyritic Mesoproterozoic granite, and the alluvium of Bullhead City of House and others (2005; fig. 4). These boulder gravels record high discharge during post-70 ka downcutting.

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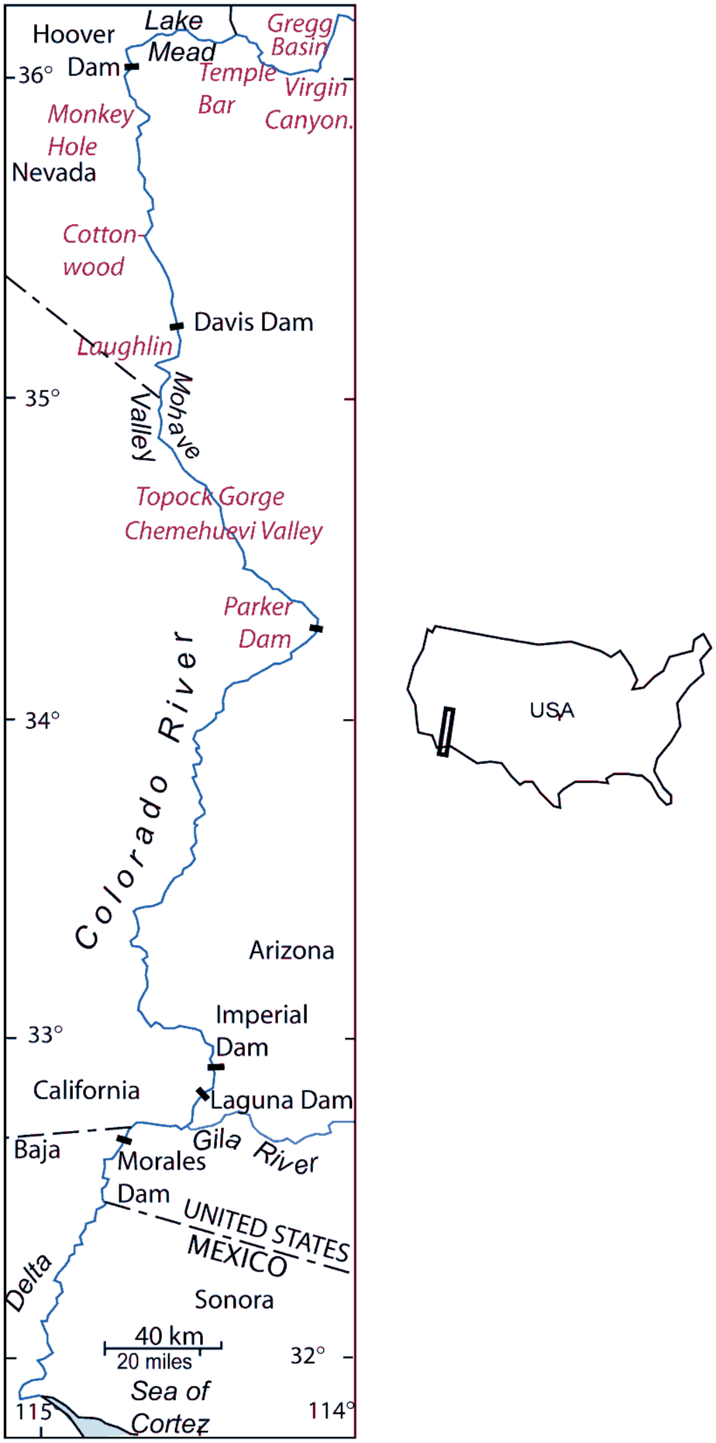

Figure 1. Localities (in red) of fluvial boulder deposits along the lower Colorado River valley, Arizona, California, and Nevada.

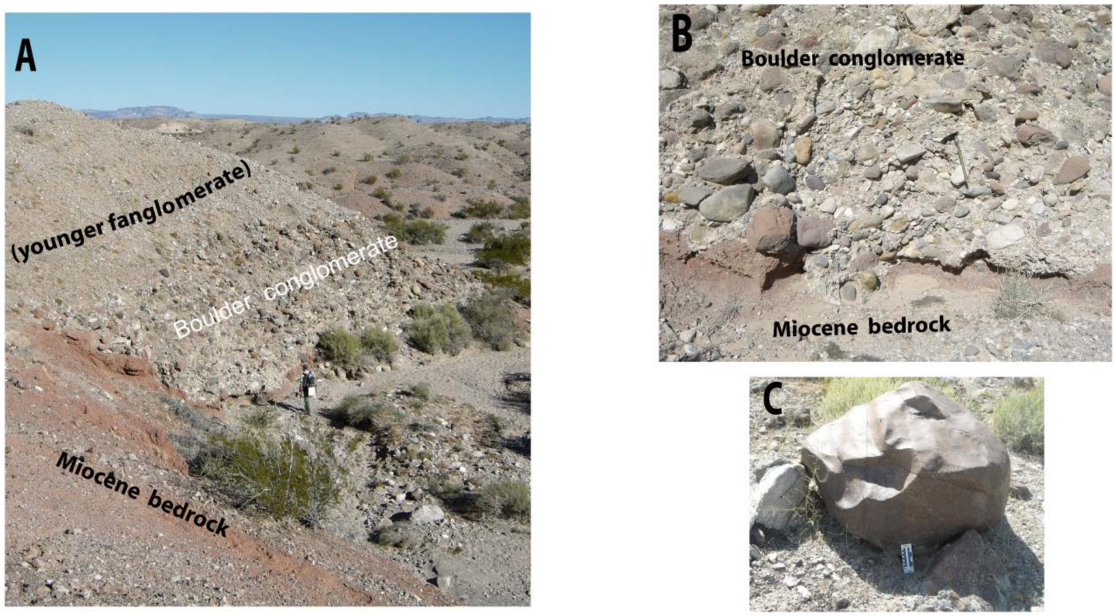

Figure 2. Cemented Pliocene(?) rounded boulder conglomerate is exposed along a 22-km-long reach of the Colorado River in Arizona and California, including Topock Gorge and parts of adjacent Mohave and Chemehuevi valleys. The deposit contains sandstone clasts derived from the alluvium of Bullhead City of House and others (2005), and contained a horse rib (pictured in Howard and Malmon, 2007). A. The conglomerate fills the thalweg of a paleovalley cut into re Miocene conglomerate bedrock in the northeast part of Topock Gorge (Arizona). A correlative boulder deposit was earlier recognized across the river in California, northwest of Topock Gorge (fig. 11 of Metzger and Loetz, 1973). The boulder unit fines upward to smaller boulders and cobbles. It appears to be a single bed 45 m thick, representing a single flood. B. Close up of the deposit shown in A. Boulders of gneiss and basalt are as large as 0.9 m across; far-traveled rounded chert clasts are as wide as 0.2 m. Imbricated boulders and large cobbles indicate current toward the camera (toward the east-southeast). The trenching tool in middle right is 0.6 m long. C. Stream-polished granite boulders, like this one, in the deposit southeast of Topock Gorge are as wide as 3 m (ruler scale is graduated in inches and centimeters).

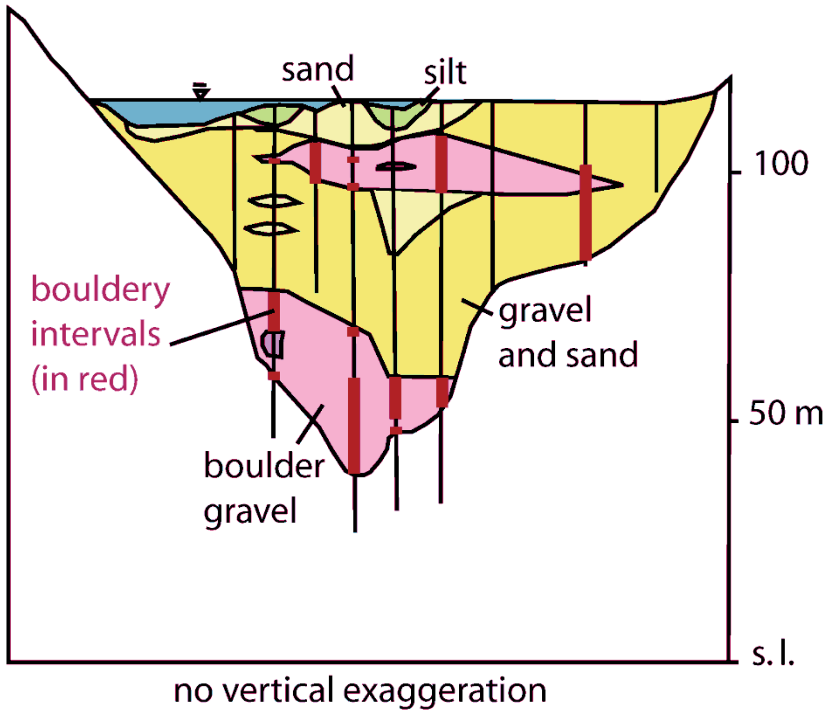

Figure 3. Cross section of boulder-bearing deposits and finer facies revealed by drill logs (vertical lines, generalized from Berkey, 1935) at the Parker Dam site near Parker, Arizona. Paleosols and cementation in the lower two-thirds of the section suggest that the lower bouldery interval is the basal part of a Pliocene or lower Pleistocene fluvial section. We interpret the upper interval as Holocene, late Pleistocene, or possibly historical.

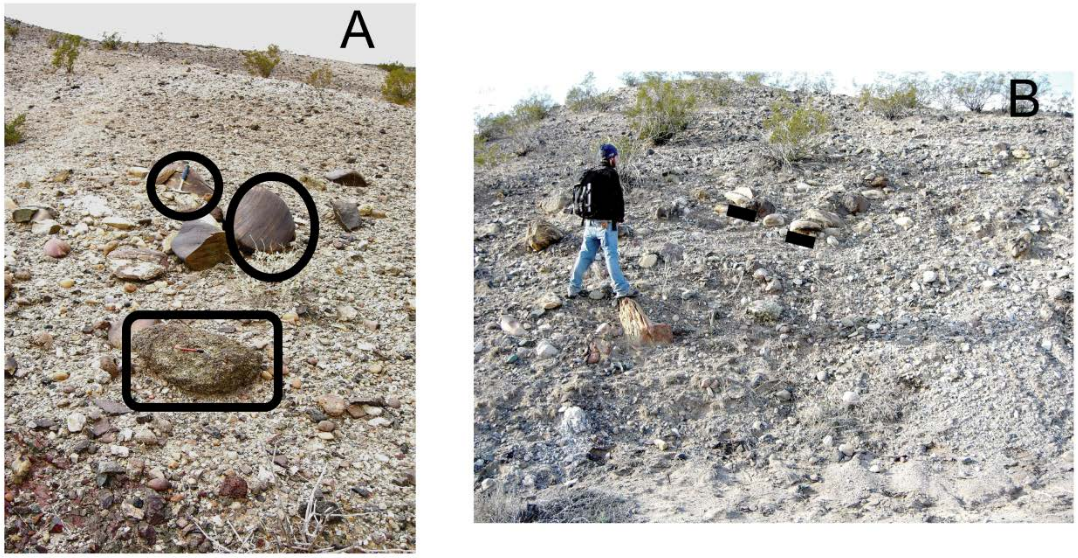

Figure 4. Boulder deposits of two late Pleistocene ages in the NW part of Topock Gorge, California, both postdate the Chemehuevi Formation of Longwell (1963). A. High terrace deposits inset into the Chemehuevi Formation include rounded large cobbles of quartzite 0.3x0.15 m across (circled, with hammer) and dark Proterozoic porphyritic granite (rectangle, with pencil for scale). The high terraces are as much as 55 m above the modern river. Nearby, the high terrace deposits include rounded boulders as large as 1.2 m, including clasts of cemented Pliocene roundstone conglomerate, gneiss, vesicular basalt, and other volcanic rocks. The clasts are likely derived from the nearby Pliocene conglomerate. B. Younger gravel that underlies a low terrace, 12 m above the modern floodplain, contains imbricated rounded boulders as long as 1 m. Downstream is to the left. Angular debris of local tributary origin caps this low terrace.

Holocene Aggradation of the Lower Colorado River in Mohave Valley, California and Arizona

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The lower Colorado River downstream between Grand Canyon, Arizona, and the U.S.-Mexico border has experienced several cycles of degradation and aggradation over its 5 m.y. history. The best dated aggradation sequence is the Holocene sediment buried under the modern floodplain.

A site 400 km upstream from the river's mouth, at Topock, near Needles, California, is being investigated as part of a remediation project for a nearby discharge of wastewater containing chromium (U.S. Department of Toxic Substances Control, 2010a, b; U.S. Department of the Interior, 2010). Holocene radiocarbon dates were obtained here from 22 wood fragments recovered from fluvial sediments in wells beneath or next to the river. Over a section 13 m thick (elevation 114–127 m), the dates decrease upward from 8.6 ka to 6 ka at an average rate of about 6 mm/year (fig. 1). Wood from levels 11 m below the 139 m modern river elevation yielded much younger ages of <500 ^{14}C yrs. B.P.

We interpret the lower part of the section at Topock to record 15 m of early to mid-Holocene aggradation. Aggradation may have been driven by eustatic sea-level rise hundreds of kilometers downstream, and (or) changes in sediment supply.

The abrupt change in the age pattern at 127 m asl indicates an unconformity, which would have removed evidence of any post-6-ka aggradation. The upper 11 m of the section, containing nearly zero-age wood, can be attributed to late Holocene scour and fill, and an additional 8 m of aggradation that occurred after the 1938 closing of Parker Dam downstream from Topock. By comparison, historical scour of ≥15 m was observed downstream in the early 1900s at Yuma, Arizona, and also was recorded upstream by a plank found buried in gravel excavated for the construction of Hoover Dam.

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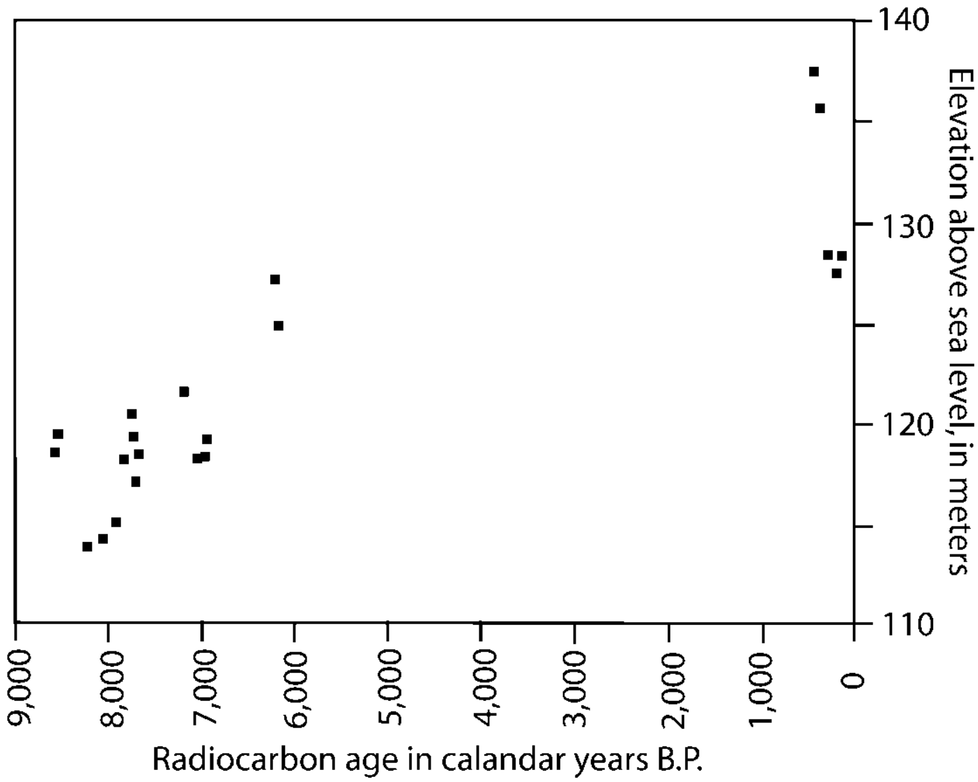

Figure 1. Ages of wood recovered from 6 core holes in fluvial sediments near Topock, near Needles, California.

Evidence from the Colorado River System for Surface Uplift of the Colorado Rockies and Western Colorado Plateau in the last 10 Ma Driven by Mantle Flow and Buoyancy

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The topography of the Earth's surface provides important information for regional and global geomorphic studies because it reflects the interplay between tectonic-associated processes of uplift and climate- associated processes of erosion. The actively deforming western U.S. Cordillera is characterized by high relief and regionally high elevation, typically exceeding 1.5 km, and invites the use of topographic analysis to further our understanding of the underlying geodynamics driving continental uplift. Intriguingly, much of the high elevation coincides with thin or attenuated continental crust, necessitating topographic support by anomalous buoyancy of the mantle. In particular, while the neighboring provinces of the Colorado Plateau and Southern Rocky Mountains are quite distinct in geology and physiography, they share recent uplift which we hypothesize to be driven by buoyant mantle.

The Colorado River system, which drains the Southern Rocky Mountains and transects the Colorado Plateau, provides an ideal natural laboratory to evaluate the relationship between the topographic character of the landscape and sub-crustal processes that potentially drive topographic uplift. The Colorado River has a double concave-up longitudinal profile with a prominent change in gradient, or knickpoint, at Lees Ferry, Arizona (fig. 1). This knickpoint separates the Lower and Upper Colorado River basins, each with different but interacting uplift histories. Detrital zircon data from the 13-6 Ma Hualapai Limestone (see Pearce and others, this volume) support previous models that the Lower Colorado River profile has evolved, and Grand Canyon has been incised, during the last 6 Ma due to drainage integration across the Kaibab uplift. We hypothesize that this integration was a response to base level fall associated with opening of the Gulf of California and mantle-driven Neogene surface uplift of the southwestern Colorado Plateau (Karlstrom and others, 2008; Van Wijk and others, 2010).

Quantitative analysis of river profiles at regional scale (fig. 2) shows that normalized river gradients in Grand Canyon are steeper than expected for its downstream position in the Colorado River system (forming the lower concavity of the double concave profile). Tributaries that join this reach of the Colorado River also exhibit increases in normalized gradient as they approach the main stem, consistent with a longlived transient response to drainage basin integration (Cook and others, 2009). In the upper Colorado River, a comparison between the upper Colorado and Green rivers shows the upper Colorado River to have steeper normalized gradients relative to the Green River. This is contrary to what might be predicted given that the Colorado River has higher historic discharge, higher incision rates over timescales ranging from 500 ka to 10 Ma, and a similar mixture of substrate lithologies (Aslan and others, 2007; Darling and others, 2008). Hence, steeper gradients in rivers of the Colorado Rockies may reflect differential rock uplift of the Colorado versus Wyoming headwater regions. This difference is also reflected in regional topographic roughness of the Grand Canyon and Colorado Rockies relative to the Colorado Plateau (fig. 3).

The observation that differential incision is balanced by fault slip in Grand Canyon (Pederson and others, 2002; Karlstrom and others, 2007; 2008) is a powerful tool to understand interplay between tectonic and geomorphic forcings. Likewise, differential incision across the Lees Ferry and other knickzones in the Colorado River system suggests they are transient incision pulses that migrate through the system in several million years. Incision rates across the Lees Ferry knickpoint in the last few million years vary from 150-175 m/Ma below, 200-500 m/Ma within, and < 100 m/Ma above this knickzone (see Darling and others, this volume), necessitating its continued upstream propagation and evolution. Additional high quality incision rate data are needed to further unravel interaction of river incision and salt tectonics (in Canyonlands and upper CR system), and should also be able to test models for migrating surface flexures related to broad epeirogenic uplifts (Moucha and others, 2009).

Different apatite fission track (AFT) and apatite He (AHe) cooling ages and denudation rates above and below the Lees Ferry knickpoint are also compatible with differential rock uplift and knickpoint transience at 10 Ma timescales (Kelley and others, this volume). AFT and AHe thermochronology in the upper Colorado River, near Rifle Colorado, include a combination of drill hole data from the MWX well and the adjacent White River uplift to give a composite age-elevation traverses from -800 m to 3200 m elevation. These data show denudation rates of 70 m/Ma from 60-40 Ma, rates of 1-5 m/Ma from 40-10 Ma, and onset of rapid exhumation at 150 m/Ma about 6-8 Ma. Similar high rates of denudation (100 m/Ma) seem to have simultaneously affected elevations ranging from 1.5 – 4 km in several places in Colorado (San Juan Mountains, Gore Range) in the last 10 Ma. This onset of rapid incision over large regions in the Colorado Rockies at 6-8 Ma is not explained by climate change at ~3.5 Ma, nor by upstream propagation of incision driven by lower Colorado River integration across the Kaibab uplift at 6 Ma, and hence is best explained by Neogene eperiogenic uplift of the Colorado Rockies.

Neogene and ongoing epeirogenic uplift of the western edge of the Colorado Plateau edge relative to the Basin and Range, and of the Colorado Rockies relative to the central Colorado Plateau are both consistent with mantle seismic tomography inversions (fig. 1; Schmandt and Humphreys, 2010). High gradient reaches (Grand Canyon and Colorado Rockies) overlie low velocity mantle suggesting that differential uplift is due to buoyancy and flow pressures exerted on the base of the lithosphere by upwelling low density mantle. The Lees Ferry knickpoint is located above a region of a sharp ~6 % Pwave velocity gradient in the upper mantle (figs. 1 and 3). Mantle-driven tectonism is also documented by the eastwards sweep of basaltic magmatism

(Wenrich and others, 1995; Roy and others, 2009) and a Neogene change from lithospheric- to asthenospheric-sourced basalts in the western Grand Canyon region (Crow and others, submitted). Presence of mantle 3He in Grand Canyon and Colorado hot springs is interpreted as the youngest tectonic signal of regional mantle-driven uplift (Crossey and others, 2009).

Further evidence for links between surface topography and lithospheric structure is shown in figure 3. Physiographic boundaries correspond to changes in Bouguer and Isostatic gravity and seismic attenuation Q (fig. 3). Lithospheric geoid anomalies of 3 to 5 meters spatially correlate with elevated long-wavelength topography along the western edge of the Colorado Plateau and the Southern Rockies in Colorado (fig. 3). The isostatic response to denudation of the Colorado River also accounts for several hundred meters of rock uplift, but is probably insufficient to account for the total magnitude of modeled and inferred uplift indicating the need to continue to refine models that can parse tectonic versus isostatic uplift components (Lazear and others, this volume). The ongoing CREST geophysical experiment (Aster and others, 2009) is providing improved seismic images that will further constrain the buoyancy structure of the lithosphere and the dynamics of sub-lithospheric processes driving surface deformation.

Evidence for ongoing uplift is supported by the conclusions of geodynamic studies that hypothesize that mantle flow continues to influence topography in the Colorado Plateau region (van Wijk and others, 2010). Our models predict small scale mantle convection, formation of lithospheric drips, and delamination of lithosphere, and translate modeled temperature structures into synthetic seismic wave velocities that match well with available tomography. Thus, a combination of mantle processes may be implicated in surface uplift. These include upper mantle buoyancy variations and mantle flow pressures (Moucha and others, 2009), perhaps combined with upwelling thermal (Yuan and Romanowicz, 2010) and/or 410-km melt-layer instabilities (Leahy and Bercovici, 2007) from the mantle transition zone. Present models indicate the importance of edge-driven convection along the western edge of the Colorado Plateau (van Wijk, and others, 2010) and lithospheric drips and return flow under the Colorado Rockies may explain the history of exhumation and uplift in these two apparently disparate provinces.

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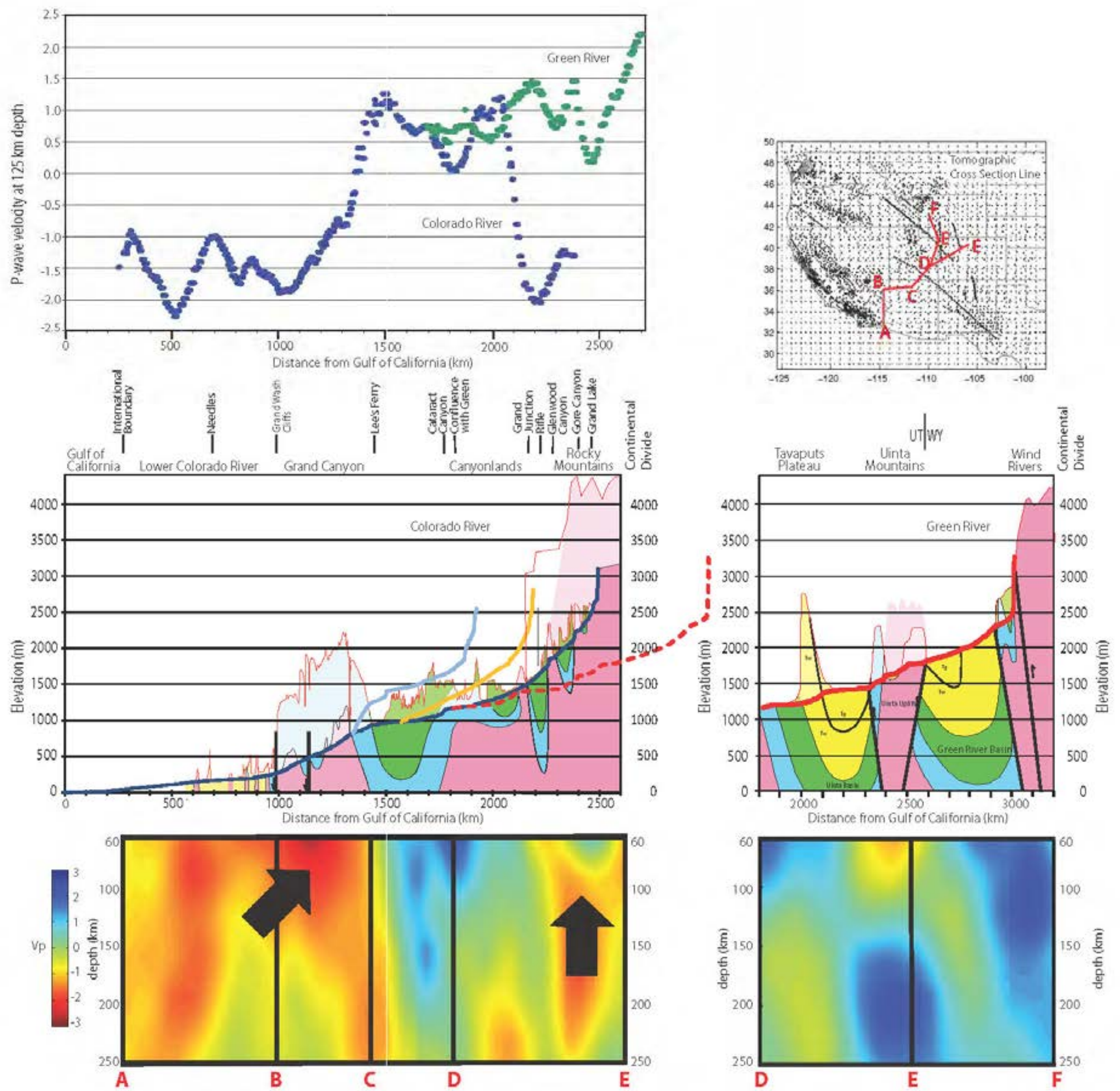

Figure 1. Longitudinal river profiles of the Colorado and Green River systems showing bedrock type and height of canyon walls. In order of decreasing erosional strength: Pink- Precambrian crystalline rocks; Blue= Paleozoic strata; Green= Mesozoic strata; Yellow= Tertiary strata. Tomographic cross sections beneath the river profiles show low velocity mantle beneath the Grand Canyon and Rocky Mountain reaches and a large velocity contrast beneath the Lees Ferry knickpoint. Buoyancy and rheology differences associated with these velocity gradients are interpreted to be driving mantle flow and ongoing surface uplift (black arrows). Upper left diagram also shows this contrast of P-wave velocity beneath the upper Colorado versus Green rivers that is interpreted to be driving Neogene uplift of the Colorado Rockies.

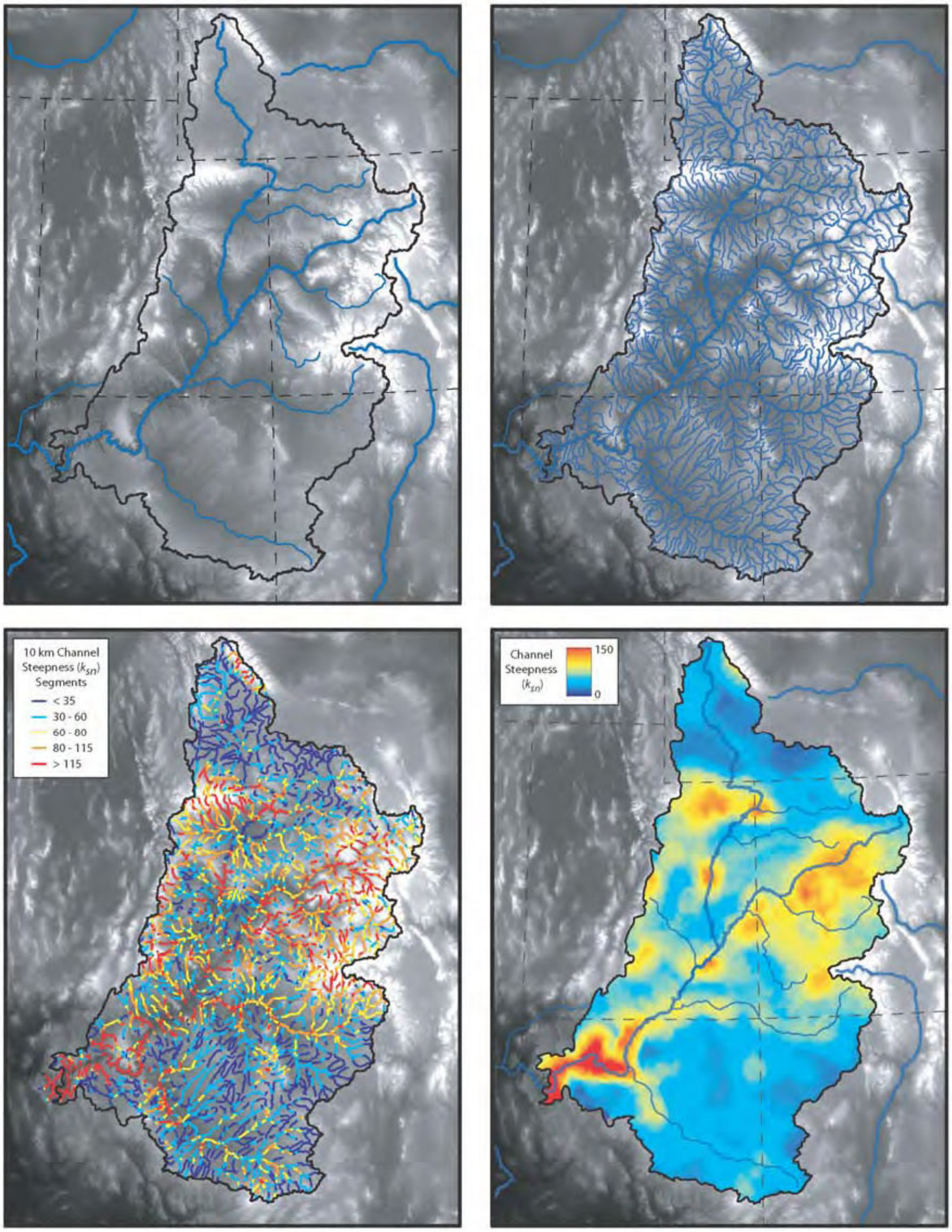

Figure 2. Summary of channel steepness (Ksn) analysis. (a) Map highlighting the major rivers and extent of the Colorado River watershed considered in our analysis. (b) Channels with drainage area >150 km2; Ksn analysis was limited to these channel reaches. (c) Individual Ksn values, calculated over 10 km long river segments (using a reference concavity of 0.45). (d) Interpolation of Ksn values. Note the broad area of high channel steepness within the upper Colorado River in comparison with the lower channel steepness values within the Green River watershed.

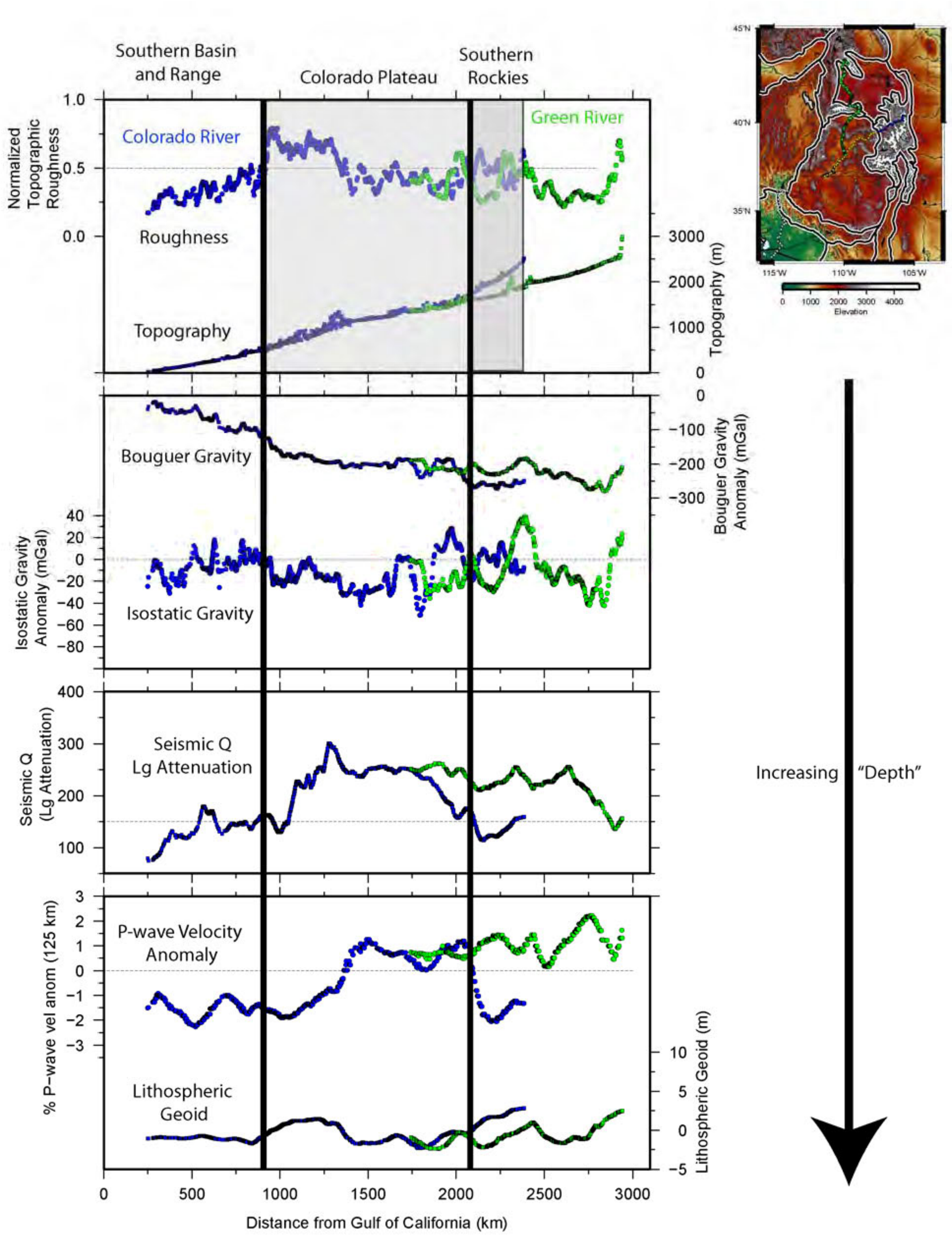

Figure 3. Relationships between topography and lithospheric parameters along the Colorado and Green river profiles.

A Summary and Evaluation of Thermochronologic Constraints on the Exhumation History of the Colorado Plateau–Rocky Mountain Region

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A regional compilation (SW to NE) of apatite fission-track (AFT) and apatite (U-Th)/He (AHe) data for the Colorado Plateau/ Rocky Mountain region shows a multi-stage history with Laramide (80–40 Ma), middle Cenozoic (35–25 Ma) and late Cenozoic (10 Ma –present) uplift/denudational components. In this paper, we examine the relative importance of each exhumation "episode" within different subregions using low temperature thermochronology that records cooling of rocks through 110–60 ºC (AFT) and 70–40 ºC (AHe). Converting these data to paleodepths and exhumation magnitude requires assumed paleogeothermal gradients. Recognizing many caveats for estimating geotherms and closure temperatures, we use 20–25 ºC/km as a likely geothermal gradient for the Plateau and Rockies during late Cretaceous to Cenozoic time. Consequently, 110 ºC = 4–5 km burial depth for AFT ages near the base of the partial annealing zone (PAZ), 60–110 º = 3–4 km depths for AFT ages within their PAZ, 40–70 ºC = 2.5– 3 km depths for AHe partial to total retention. Rates of exhumation are approximated by regressing age-elevation traverses, with no assumption about evolving geotherms or variable closure temperatures, but with the assumption of steady geotherms and semi-uniform closure temperatures over the time period. Figure 1 shows the different subregions discussed below keyed to the numbered paragraphs. Figure 2 shows AFT age-elevation traverses in the Grand Canyon region, and figure 3 shows selected AFT age-elevation traverses in the Upper Colorado River basin region.

1) Laramide AFT cooling ages of 90 to 60 Ma in the Mogollon highlands (Bryant and others, 1991; Foster and others, 1993; Fitzgerald and others, 2009) along the southwestern margin of the Colorado Plateau suggest stripping of 2–4 km of sedimentary cover in the late Cretaceous and early Tertiary, and unroofing of Precambrian basement before 50 Ma (Rim gravels) to 18 Ma (Peach Springs Tuff). Paleocanyons of km-depth-scale flowed NE off the highlands in the Salt River and Peach Springs canyon areas during the late Eocene time. Drainages reversed and flowed southwards in late Miocene (Young, 2001, Potochnik and Faulds, 1998).

2) A similar Laramide cooling history is documented in the Virgin Mountains (Quigley and others, 2010; Fitzgerald and others, 1991, 2009) by ca. 90 Ma ages that are still locally preserved near the edges of the western Colorado Plateau. But the dominant thermochronologic signature is of rapid tectonic denudation via normal faulting of 3–5 km sedimentary cover at 17–15 Ma. The North Virgin Mountains show a transition from slow (3–4 ºC/m.y.) to rapid (12–25 ºC/m.y. or

500–1000 m/m.y.) cooling and denudation at 22–17 Ma and evidence for a pre-17 Ma burial depth of ~ 5 km (Quigley and others, 2010). AFT and AHe, and zircon (U-Th)/He data from the Gold Butte block also show rapid cooling of basement rocks from 110 to 50 ºC at 17 to 15 Ma. Concordance of AFT and AHe dates suggest very fast cooling due to high tectonic denudation rates (~2000 m/m.y.). The 110 ºC isotherm (top of PAZ) at 17 Ma was 200–300 m below the Great Unconformity (Reiners and others, 2000, Fitzgerald and others, 1991, 2009; Quigley and others, 2010), compatible with 3–4 km of Phanerozoic sedimentary cover in the South Virgin Mountains. The thermochronologic data suggest northwards thickening of Phanerozoic cover before 17 Ma and a land surface that was ~ 2 km higher than the Kaibab Limestone. This precludes a precursor >17 Ma Grand Canyon cut into Kaibab Limestone in this location (Wernicke, 2009).

3) Western Grand Canyon AFT dates range from 114–50 Ma and AHe dates are 10–20 Ma younger for the same rocks (Lee, 2007; Flowers and others, 2007, 2008). AFT track lengths of 13 µm suggest moderate cooling rates between 114–50 Ma. A juxtaposition of 80–114 Ma AFT ages to the west (fig. 2) and 50–70 Ma to the east across the Hurricane fault indicates differential cooling across this fault system. AHe ages that are 50–80 Ma indicate that the average paleosurface was <2 km above the Kaibab Limestone, although NE-flowing rivers had cut km-deep canyons along the Hurricane fault, locally down below the Redwall Limestone. By 10 Ma, basalt flowed across a topographic surface that had developed on the Kaibab Limestone.

4) Eastern Grand Canyon AFT ages show a wider range of ages (80–30 Ma), with similar abrupt juxtapositions across Laramide faults and monoclines of the East Kaibab uplift (Naeser and others, 2001, Kelley and others, 2001). Age-elevation traverses show very rapid cooling Laramide cooling 61 to 66 Ma (Dumitru and others, 1994; Kaibab, fig. 2). More protracted Laramide cooling (79–49 Ma) is recorded in AFT data from the Tanner Trail to the east of the Kaibab traverse (fig. 2). AHe dates on basement rocks range from 23 to 55 and suggest the average paleosurface was 1–2 km above the Kaibab Limestone (Flowers and others, 2007, 2008). Selected AHe (and AFT) ages from the rim and bottom of the canyon that yield similar ages have been used to infer paleotopography (Wernicke, 2009), but geometry of ancestral paleocanyons and scarps is not well constrained by existing data. Thermochronologic evidence for a pre-6 Ma paleo “Grand Canyon” (Wernicke, 2009) is not supported by our analysis. Data from the Grand Canyon region do indicate km-scale mid-Tertiary topography on a surface that was 1–2 km above the modern Kaibab surface. This paleotopography involved NE flowing paleocanyons and N-retreating bedrock scarps. Various combinations of canyons, monoclines, and scarps can explain differential cooling, but the geometry of the now-eroded surface remains poorly constrained. We expect that additional thermochronology can refine the geometry of paleocanyons cut by NE-flowing rivers on now-eroded stratigraphic units 1–2 km above the Kaibab Limestone, such canyons should not be called “Grand Canyon” because of the confusion factor for the visiting public.

5) The monoclines in the Lake Powell to Canyonlands area have Laramide and >100 Ma AFT ages suggestive of variably tilted Laramide PAZs. Surficial and drillhole data from the Monument monocline in Utah show that the landscape in the central Plateau was stable during middle Cenozoic time and was rapidly exhumed at about 6 to 7 Ma (Hoffman, 2009). These dates may overlap, or just pre-date, the timing of the integration of the Colorado River. An

important research question that remains to be addressed is whether this rapid denudation was driven by drainage integration or by Neogene uplift.

6) The Proterozoic basement of southwestern Colorado records a complicated <110°C thermal history related to Laramide exhumation, mid-Cenozoic heating, and Miocene cooling. AFT ages decrease from ~180 Ma (in the AFT PAZ) to 54–67 Ma to 40 Ma along an east to west traverse through Proterozoic rocks along the Gunnison River valley. Samples from Proterozoic basement of the Uncompahgre uplift in southwestern Colorado have AFT ages of 23–38 Ma, with mean track lengths of 12.7 to 12.9 µm, indicative of moderate cooling rates for this shallowly buried highland.

7) The White River uplift area along the Colorado River of southwestern Colorado has AFT age-elevation data that record slow exhumation rates of 30 to 40 m/m.y. from 63 to 34 Ma (fig. 2; Naeser and others, 2002). Nearby, AFT data from shallow levels of the MWX well in the valley of the Colorado River near Rifle, Colorado record the development of a PAZ that formed during a period of landscape stability on the Plateau 25 to 10 Ma, then a sharp increase in cooling rate to about 100 m/m.y. since ca. 6 to 8 Ma (fig. 2). The cooling could be related to the incision of the Colorado River (Kelley and Blackwell, 1990), but appears to slightly pre-date integration of the Colorado River across the Kaibab uplift. Alternatively, these data may indicate exhumation related to surface uplift in the Colorado Rockies. Present incision rates are about 150 m/m.y.. This key area shows that the Colorado River system may have started rapidly incising at 6 to 8 Ma, before the upper Colorado River system could have felt downstream effects from 6 Ma river integration.

8) AFT from the Gore Range (Naeser and others, 2002) show spatially variable exhumation initiation times and rates from three age-elevation traverses: 81 m/m.y. from 20–5 Ma in the E Gore Range, 91 m/m.y. from 25–13 Ma on Mt Powell and 43 m/m.y. from 35 to 20 Ma in the western Gore Range. Taken together, the data are consistent with semi-steady denudation from 32 to 5 Ma, perhaps in response to uplift associated with the Rio Grande rift (Naeser and others 2002), and perhaps in response to regional epeirogenic uplift above the Aspen anomaly (Karlstrom and others, 2005).

9) The Sawatch Range has AFT ages from 9 to 52 Ma (Bryant and Naeser, 1980, Kelley and others, 1992; Feldman, 2010) and has a complicated Laramide plutonic emplacement and subsequent rift-flank uplift history. An episode of ca. 15 to 23 Ma cooling is recorded in the Mount Princeton and Huron Peak plutons of Oligocene age. Snowmass Mountain, a 35 Ma stock in the western Elk Mountains, yields AFT cooling ages of 21 to 29 Ma with denudation rates of about 75 m/m.y. (fig. 3).

10) AFT ages for Proterozoic rocks in the Needle Mountains are 8 to 14 Ma in Chicago Basin and 13 to 30 Ma in the Animas River valley; the ages correlate well with elevation. AHe ages for plutonic rocks of the southeastern San Juan volcanic field (20 to 39 Ma) are similar to emplacement ages, indicating rapid cooling of shallowly emplaced plutons. In contrast, the AFT and AHe data from plutons in the NW San Juan volcanic field suggest a more protracted cooling history. For example, data for 26.6 Ma Sultan Mountain Stock near Silverton reveal that it cooled below 110°C at ~23 Ma and below 70°C at ~10 Ma. The combined data indicate that a

1–3 km volcanic cover on the San Juan Mountains was stripped differentially starting about 20 Ma, with an important onset of rapid cooling in the last 10 Ma.

11) The Uinta Basin has been slowly exhuming since Oligocene time, based on AFT data derived from cores in three different drillholes.

Thermochronology alone does not constrain rock or surface uplift, but this paper shows that "relative thermochronology" (comparison of subregion to subregion) provides important information about differential uplift patterns. Age elevation traverses and drillhole data in different regions show high exhumation in areas denuded by tectonic extension, but erosional exhumation rates tend to range from tens to 100–200 m/m.y. across the region. Differences in age-equivalent denudation rates require differential rock uplift (across faults and due to differential eperiogenic uplift).

Thermochronologic data in the Colorado Plateau-Rocky Mountain region record a denudation history that reflects three main exhumation stages from 80 Ma to present. Some AFT data record Laramide (90–70 Ma) cooling below 110 °C in uplifted blocks. 70–40 Ma AFT and AHe dates reflect differential, but overall moderate, cooling as Mesozoic strata were differentially stripped from the eroding structural topography of the post-Laramide landscape (e.g. Grand Staircase and Rockies). 35 to 25 Ma AFT and AHe ages dominate the southern Rockies due to cooling and exhumation associated with large volcanic fields (e.g San Juans). The most exciting result of our regional analysis is the evidence for <10 Ma onset of rapid exhumation in the Upper Colorado River basin and Rocky Mountains that may, in part, be tied to Neogene uplift of the Colorado Rockies.

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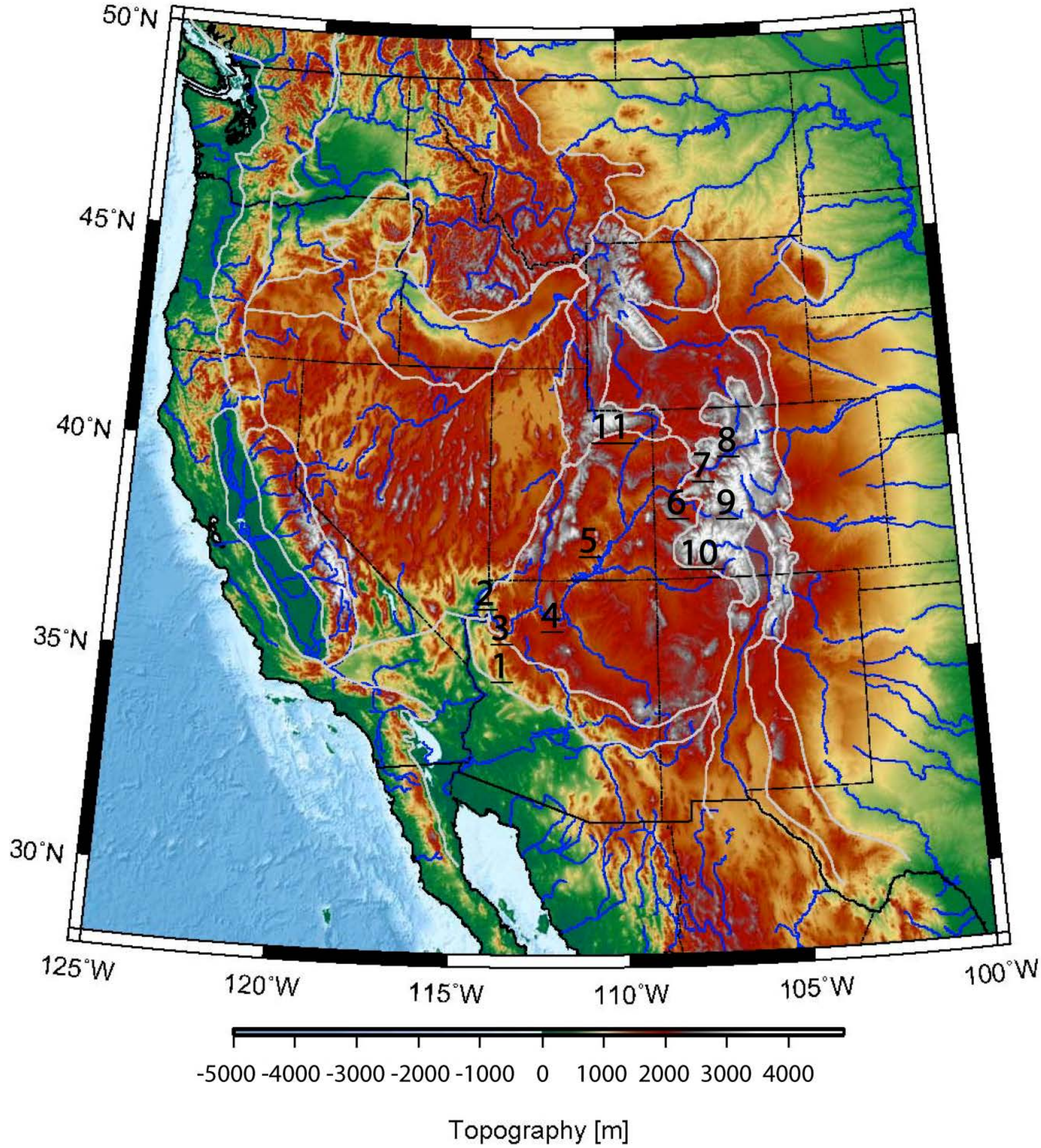

Figure 1. Relief map of the western United States. The numbered areas are keyed to the numbered paragraphs in the text.

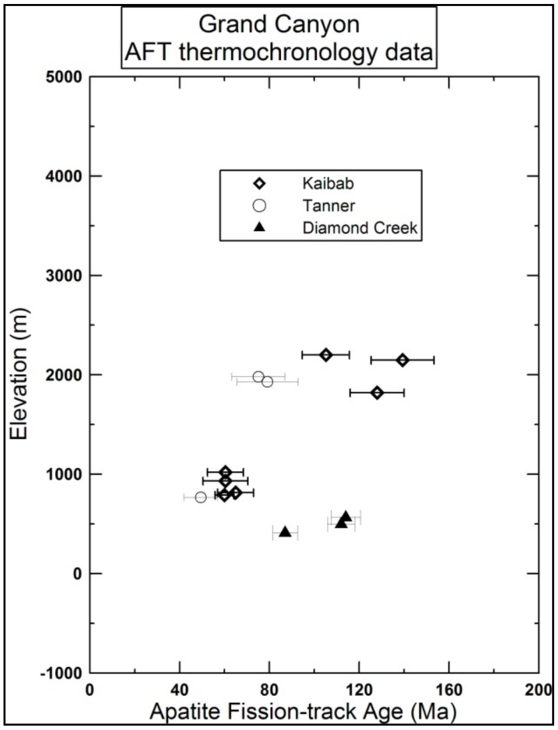

Figure 2. AFT age-elevation traverses in the Grand Canyon. Kaibab Trail data are from Dumitru and others (1994) and Tanner Trail data are from Kelley and others (2001).

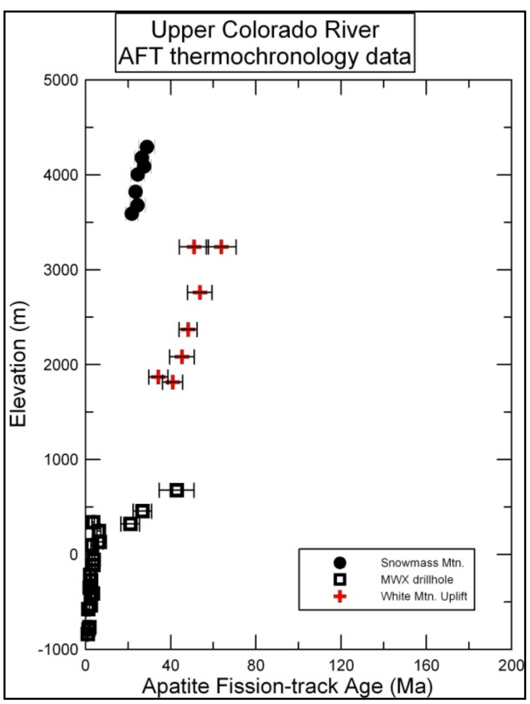

Figure 3. Selected AFT age-elevation traverses in the upper Colorado River basin.

Detrital Zircon Record of Colorado River Integration into the Salton Trough

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The Colorado River is a youthful, unequilibrated continental drainage system, the base-level for which was established rather abruptly only 5–6 million years ago in conjunction with Gulf of California rifting and establishment of the modern river course through the western Grand Canyon and lower Colorado River region in the southwestern United States (Lucchitta, 1972, 1989; Gastil and others, 1996; Howard and Bohannon, 2001; House and others 2005, 2008; Dorsey and others, 2007; McDougall, 2008; Spencer and others, 2008). A Colorado River source of Pliocene Imperial Group deltaic sediment in the Salton Trough is strongly supported by sandstone petrology and the presence of reworked Late Cretaceous foraminifera derived from the Mancos Shale of the Colorado Plateau (Lucchitta, 1972; Buising, 1990).

Laser ablation ICPMS detrital zircon U-Pb analyses (~3,000) from ~40 samples provide insight into drainage-basin evolution, including the cause, timing, and consequences of modern river integration. Samples encompass: (1) the modern Colorado River delta; (2) major tributaries including the Green, “Grand”, San Juan, Little Colorado, Virgin and Gila rivers; (3) late Miocene to Pliocene sediments along the lower Colorado River; (4) late Miocene to Pleistocene deltaic and fluvial sediments of the Imperial and Palm Spring Formation in the western Salton Trough; and (5) late Miocene-early Pliocene Bidahochi Formation of eastern Arizona.

Data from modern Colorado River delta sands (fig. 1) and exhumed Colorado River deltaic deposits (~5.3–1.0 Ma) along the western Salton Trough (fig. 2) yield statistically indistinguishable detrital zircon age distributions that indicate little or no evolution in the detrital-zircon age spectra through time. Sources for the dominant peaks include local 1.7 Ga and less common 1.4 Ga basement, major inputs from reworked Colorado Plateau strata, including Permian and Mesozoic erg deposits that provide Grenville and early Paleozoic zircons (Dickinson and Gehrels, 2009), and Mesozoic basement and 12–25 Ma igneous rocks downstream from the Grand Canyon. Archean zircons were likely sourced from the Wyoming province through the Green River and by reworking from the ergs and other sediments partly derived from Wyoming.

The basic Colorado River “delta DZ reference signal” is established far upstream in the modern Colorado River system by the Green, “Grand”, and San Juan tributaries (fig. 3). Detrital zircon contributions farther downstream from the distinctive Little Colorado, Virgin and Gila drainage basins have little or no effect on the well-established detrital-zircon spectra already being carried along by the river.

Detrital-zircon age patterns from the Bidahochi Formation and the Little Colorado River match one another closely, but are both clearly distinguished from the Colorado River reference by a relative abundance of early Mesozoic grains. This result is consistent with Bidahochi paleodrainages as precursors to the modern Little Colorado River, as opposed to representing fluvial input from an ancestral Colorado River that flowed southeast into the Bidahochi basin.

A Two-Stage Model for Gila and Colorado River Integration into the Salton Trough

The impressive consistency of detrital-zircon ages of Colorado River sediments through the Salton Trough and along the lower Colorado corridor suggests that the modern drainage basin was integrated at the inception of sediment delivery through the western Grand Canyon into the Grand Wash area, consistent with a lake-spillover hypothesis for initiation of the lower Colorado River (for example, Dorsey and others 2007).

However, there is an apparent paucity of Grenville age and 400-600 Ma grains in basal "Colorado-River-derived" Salton Trough sediments in the Split Mountain gorge section; this feature more closely matches the detrital-zircon age distribution of the modern Gila River as opposed to the Colorado River delta reference signature. The two basal SaltonTrough samples with extraregional detrital-zircon signatures are from sandy thick-bedded turbidites of the Wind Caves Member of the Latrania Formation that are precisely dated between 5.33 and 5.24 Ma based on fossils and magnetostratigraphy (Dorsey and others, 2007). The Wind Caves strata, in turn, are overlain by prograding muddy pro-delta deposits of the Mud Hills Member of the Deguynos Formation associated with an abrupt shift in paleobathymetry from middle bathyal to upper bathal/neritic water depths. The slight contrast in detrital-zircon patterns for Wind Caves versus Deguynos and younger strata could conceivably be explained by the Gila River feeding into the Salton Trough and depositing the Wind Caves strata *prior* to the integration of the Colorado River through the western Grand Canyon.

In summary a serial two-stage model of river integration could explain:

- A potentially more "Gila-like" detrital-zircon signature of basal Wind Caves Member strata.
- The apparent age paradox of the 5.33–5.24 Ma age for the arrival of Colorado River sediments in the Split Mountain Gorge section versus the 4.8 Ma tephra correlation age for the Bouse Formation below Grand Wash.
- The presence of coarse, thick-bedded turbidites at the base of the Split Mountain section (Wind Caves Member) overlain by prograding pro-delta deposits.
- The absence of reworked Late Cretaceous foraminifera from the Mancos Shale in the Wind Caves Member (Merriam and Bandy, 1965), because these reworked foraminifera were likely derived from outcrops to the east of the Kaibab upwarp in the western Grand Canyon, but not from the Gila River drainage area.

This two-stage integration scenario does not conflict with the story of lake spillover explaining the initial lower Colorado River downstream of Grand Canyon (House and others, 2005; Spencer and others 2008), although it adds the additional complication that the Gila River fortuitously and independently preceded the Bouse Formation and the Colorado River integration by at least a half million years. This interpretation predicts that Mancos Shale foraminifera do not occur beneath the basal Mud Hills Member of the Imperial Group. This hypothesis also suggests the possibility that a proto-Gila River may have been feeding into the Salton Trough *much* earlier prior to integration of the river through the western Grand Canyon, and that these sediments may be preserved as the middle to late Miocene ~12–6 Ma (mostly buried) "proto-Gulf" sediments best preserved in exploratory wells for oil in the northern part of the Gulf of California (Helenes and others, 2009). The two-stage model presented here requires more investigation and must be tempered by the fact that we are using the modern Gila River detrital-age zircon age distribution as a proxy for the detrital zircon age distribution of this drainage basin 5 million years ago.

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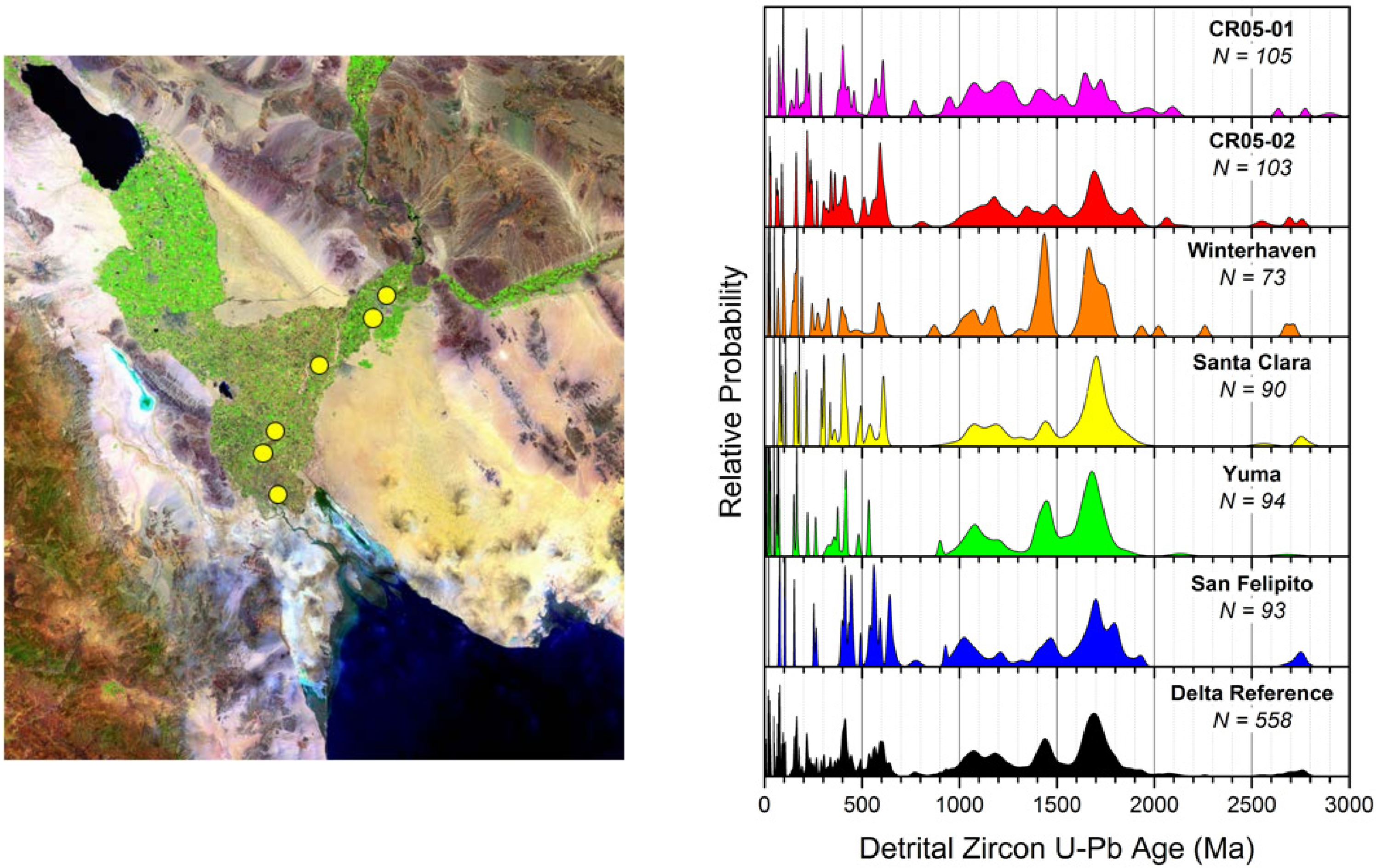

Figure 1. Landsat image of Colorado River delta showing sampling localities of modern delta sand for detrital zircon U-Pb analysis. Age specta of detrital zircon from individual samples are shown to the right illustrating similarity of results from different samples, which are combined to form the Delta Reference.

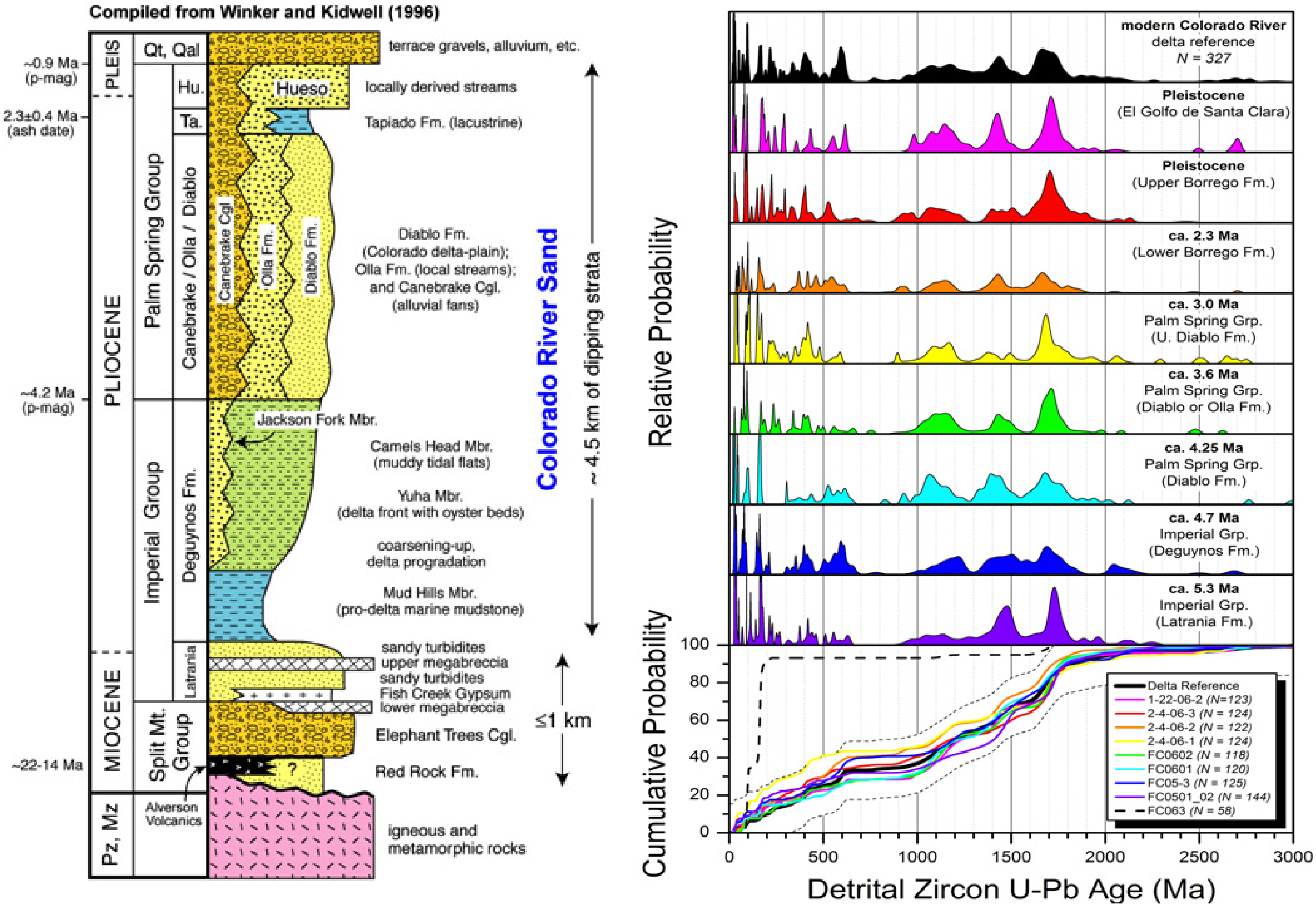

Figure 2. Stratigraphy of the Split Mtn-Vallecitos basin along the western Salton Trough showing abrupt arrival of Colorado River delta sediments to the basin near the Miocene-Pliocene boundary. To the right are DZ results from this section and other Salton Trough localities depicting the historical record of delta sediment input over the past ~5.3 Ma; modern Colorado River delta reference for comparison. Distinct sample FC063 in the cumulative probability summary is from eastern Peninsular Ranges-derived sandstone in the Latrania Fm just beneath basal Colorado River sandstone.

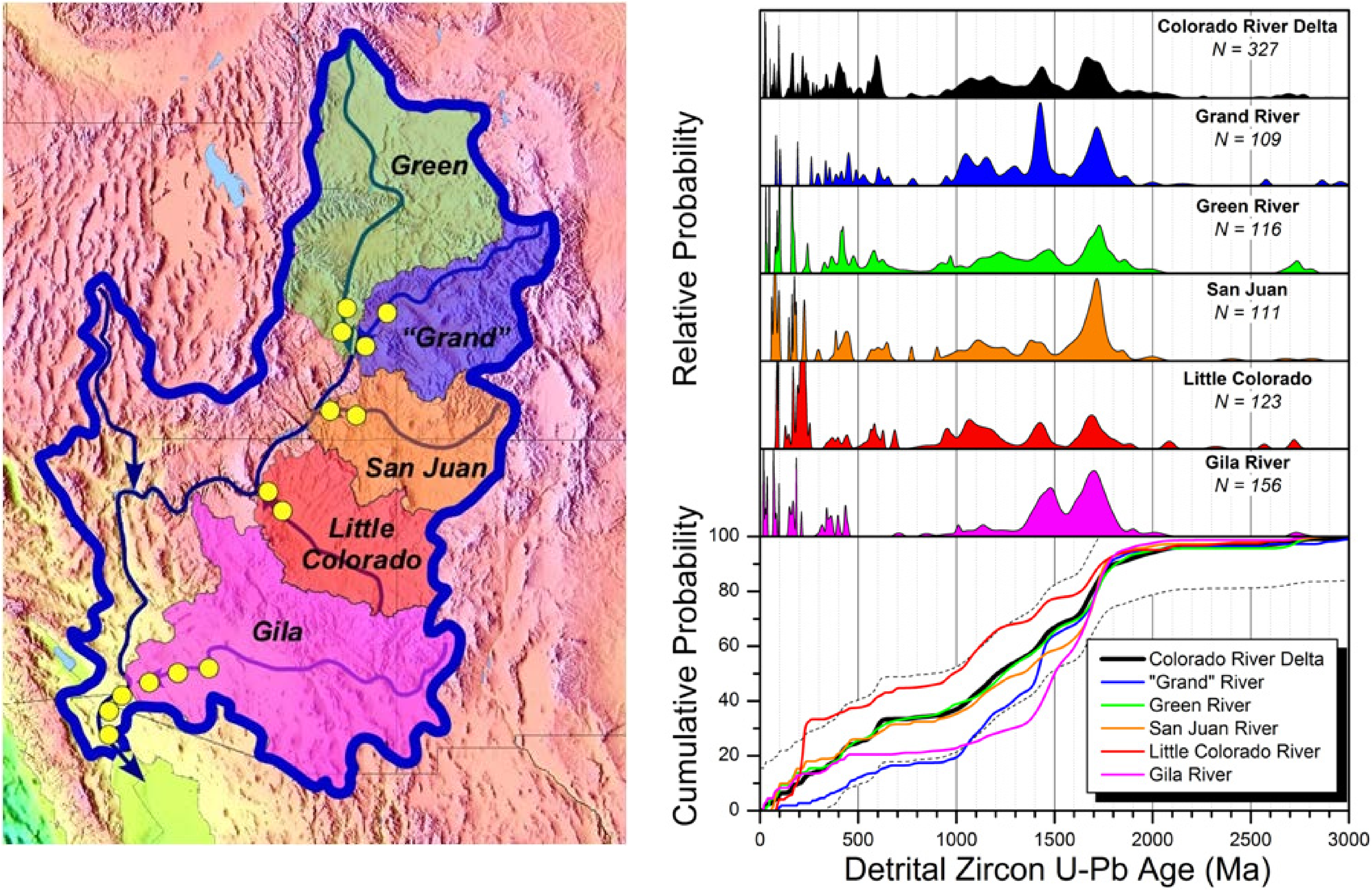

Figure 3. Colorado River drainage sampling of modern river sands. The Colorado River delta detrital zircon spectra is established in the upper Colorado River drainage basins by the Green, Grand, and San Juan river systems. This signature is little affected downstream by distinctive inputs from the Little Colorado, Virgin, and Gila rivers.

Unroofing and Incision of the Grand Canyon Region as Constrained through Low-Temperature Thermochronology

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How the Colorado River came to follow its modern course has been a question that has gone without a conclusive answer for decades. The unlikely course of the Colorado River in the Grand Canyon region, however, is undoubtedly linked to the unroofing and landscape evolution patterns. Since the latest Cretaceous, more than two kilometers of sedimentary accumulation have been removed. The accurate description of how and when this sedimentary stack was removed is a crucial step in describing how the Colorado River established its modern course and how Grand Canyon came to be.

This study considers the data collected from a subset of samples in the Grand Canyon region. In addition to multiple vertical transects and core samples from the surrounding plateau, a lateral transect was collected along river level from Lee's Ferry to the exit point of the Colorado River from the Colorado Plateau. These samples were initially analyzed for apatite fission-track ages, and more recently for apatite (U-Th)/He ages. In this way, we obtain both ages from the same rock sample. Apatite fission-track results have been partially published (see Kelley and others, 2000). In this study, we present complementary and unpublished (U-Th)/He data for 14 of the samples that yielded sufficient apatite. Figure 1 shows the geographic distribution of the river samples used in this study.

Fission-track ages range from 28.4 Ma in the east to a maximum of 73.5 Ma in the western samples (fig. 2). One sample situated on the east flank of the Kaibab Uplift, GC49, yields the oldest observed age of 78.3 Ma owing to its residence higher in the stratigraphic position. (U-Th)/He ages range from 7.7 Ma in the easternmost sample to 82.5 Ma in the western samples. Ages generally are older atop the Kaibab Uplift and in the west, with a clear younging trend east towards Lee's Ferry. Age/eU (effective uranium concentration) correlations are observed in the data collected from sedimentary rocks, as differing source terrains contribute grains with varying uranium and thorium concentrations. Samples collected from crystalline rocks do not show such variability due to the intra-sample homogeneity of uranium concentrations. Because diffusion of helium in apatite grains is partially dependent on alpha damage, grouping individual analysis by eU concentrations allows more constrained modeling of viable thermal histories (for more, see Shuster and others, 2006 and Flowers and others, 2009). As a result, samples collected from sedimentary rocks have individual analysis grouped by characteristic eU (as shown in fig 2.).

To obtain accurate thermal histories for each of the samples, fission-track and (U-Th)/He age data were modeled using the inverse modeling software package, HeFTy (Ketcham, 2005). Viable thermal history envelopes are identified by the Monte Carlo simulation method, where random thermal histories are generated, and the resulting modeled ages are compared to observed age spectra. Model fission-track and (U-Th)/He ages are calculated using the most recent parameters available (Ketcham and others, 2007; Flowers er al., 2009). Figure 3 shows selected (representative) thermal histories from samples collected along the river. For each sample, a viable temperate range is extracted at given time slices of the modeled best-fit, time-temperature history. Assuming a geothermal gradient (here, 25°C/km,) it is possible to calculate an unroofing history. When plotted in projected cross-section, both localized and regional exhumation patterns are easily distinguished (see fig. 4 as discussed below).

The results of this study clearly identify three major cooling events in the area of the Grand Canyon. The earliest cooling event occurs in the late Cretaceous to early Paleocene and is observed in the westernmost samples, as well as in those on the Kaibab Uplift. We attribute this cooling event to late-Laramide uplift of the Kaibab and Toroweap monoclines. Secondly, a major cooling episode is constrained to the Miocene and is observed in all Grand Canyon area samples. This widespread cooling event indicates the removal of a vast majority of the remaining Mesozoic sediment and also indicates incision of a paleocanyon through the Kaibab limestone and spatially coincident with the modern eastern Grand Canyon. However, all samples east of the Kaibab Uplift indicate that at significant thickness of Mesozoic sediment still occupied that region. The latest cooling event occurs post-Miocene and is observed in the easternmost samples near Lee's Ferry. During this cooling event, the 2-km thick section of Mesozoic overburden is almost completely removed. This event most likely reflects knickpoint migration as a result of the ~6 Ma drainage integration and base-level fall of the Colorado River system.

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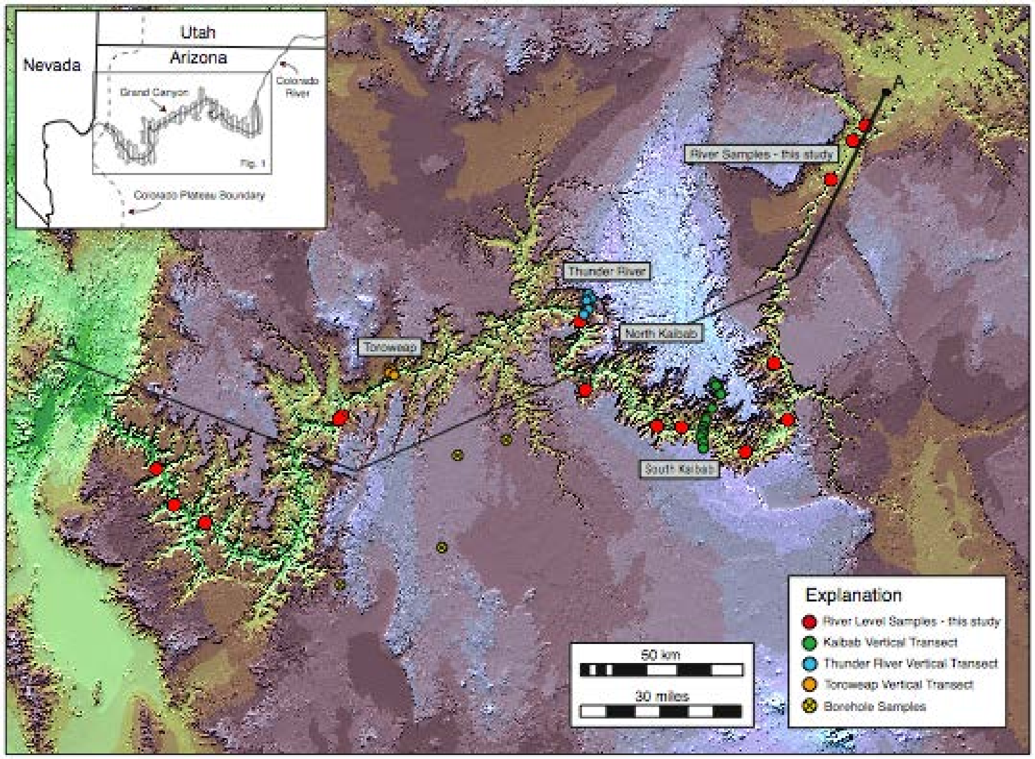

Figure 1. A painted relief map of the Grand Canyon region of the Colorado Plateau (produced from 1 arcsec NED digital elevation model, obtained from *http://seamless.usgs.gov*). The map shows the major physiographic features of the region, including Grand Canyon, the Kaibab Uplift, and the Grand Wash Cliffs. Collected sample sets are shown as colored circles. This study uses the river samples (red circles). The black line shows the major trend of the Colorado River and is used to project a sample's distance downstream.

Age Profile

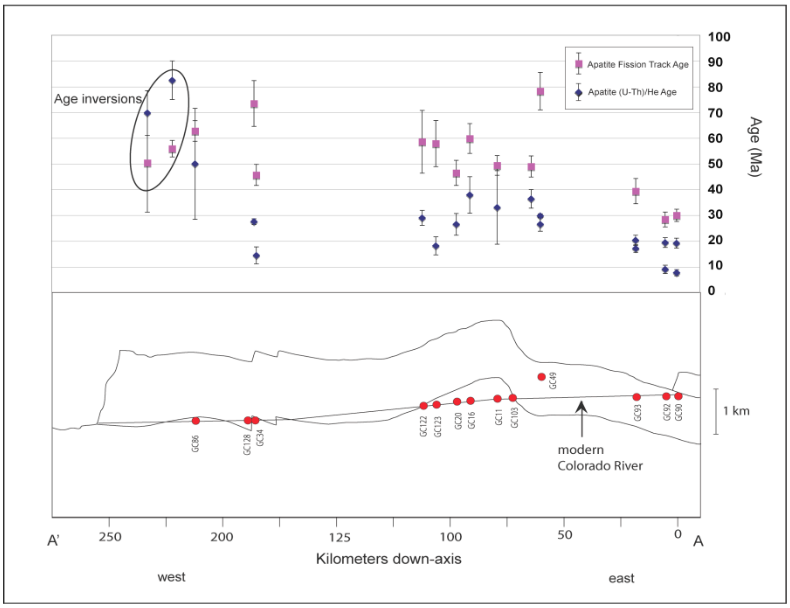

Figure 2. An age profile plot of apatite fission track and apatite (U-Th)/He age pairs. Distance down-axis is the distance downstream from Lee's Ferry as projected to the major trend of the Colorado River, Arizona (see fig. 1).

Selected Thermal Histories

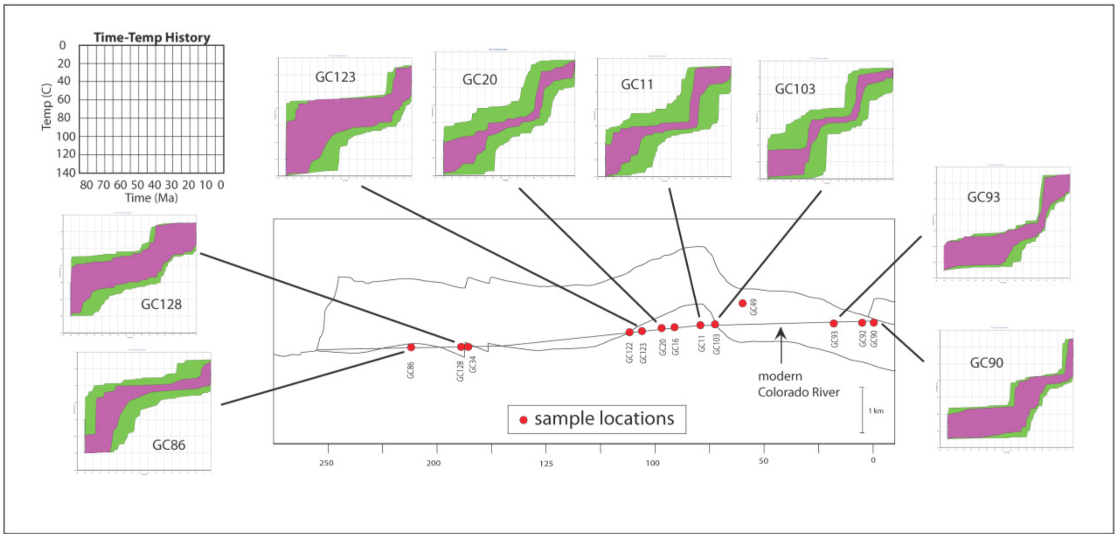

Figure 3. A schematic cross-section of the Grand Canyon region, Arizona, showing representative modeled thermal histories using apatite fission track and apatite (U-Th)/He age data. The thermal histories are produced using the inverse modeling software package, HeFTy. Cooling events can be identified from near-vertical paths in the time temperature plots. The thermal histories produced and displayed here are used to calculate unroofing patterns (see next figure). Best-fit thermal windows are displayed in purple. Acceptable-fit thermal windows are displayed green.

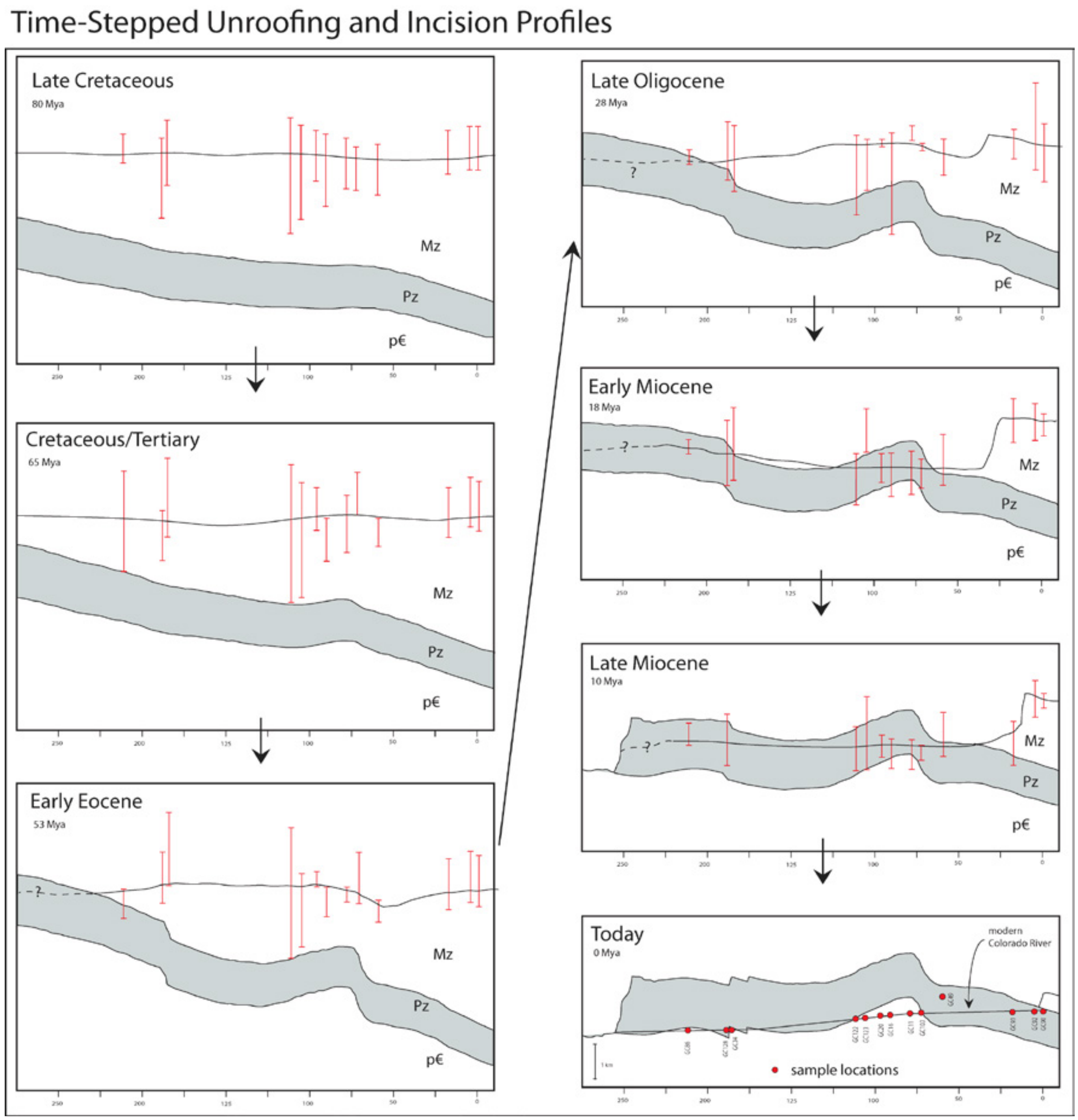

Figure 4. Time stepped unroofing profiles based on the modeled thermal histories for each sample, Grand Canyon region, Arizona. Time slices are extracted from the continuous thermal histories to show the interpreted landscape at periods before and after major cooling events. The vertical red bars represent the possible ranges in Mesozoic overburden thickness for the given time step upon which the interpreted Mesozoic surface is based. The assumed geothermal gradient is 25°C/km. Three major cooling events are identified: (1) early Paleocene; (2) Miocene; and (3) Pliocene to present.

Syntectonic Deposition and Paleohydrology of the Spring-Fed Hualapai Limestone and Implications for the 6-5 Ma Integration of the Colorado River System through the Grand Canyon

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The Hualapai Limestone is in the eastern Lake Mead region, at the western edge of the Colorado Plateau in Arizona and Nevada. It is a key unit for understanding the integration of the Colorado River from the Colorado Plateau to the Basin and Range province and the Neogene topography of the Grand Canyon region because it provides the best available middle and late Miocene (12 – 6 Ma) sedimentary record at the mouth of the Grand Canyon. The goal of this study is to better understand the early evolution of the Colorado River system by examining variations of the sedimentary and tectonic records from basin to basin. We accomplished this by refining the structural, stratigraphic, and sedimentologic data with new geochemical data, detrital zircon data, and tephrochronology.

Structural studies indicate the presence of listric faults, half grabens, and 5 – 11 km depth to detachment for the faults of the eastern Lake Mead region. Measured stratigraphic sections highlight west-to-east thickness variations for the Hualapai Limestone in two of the four basins: Grand Wash trough (10 – 212 m) and Gregg Basin (12.5 – 120 m). Gregg Basin and Grand Wash trough have east-thickening wedge-shaped, half graben geometries with depocenters adjacent to the Wheeler and Grand Wash faults, respectively. Counter to models that postulate that the Muddy Creek Formation and Hualapai Limestone are post-tectonic (Bohannon, 1984; Faulds and others, 2001), these basin geometries indicate that fault slip provided accommodation space for syntectonic deposition of the 12 – 6 Ma Hualapai Limestone in fault-controlled basins. Soft sediment deformation, seen as contorted bedding, dewatering features, and low angle unconformities also indicate syntectonic deposition.

A sedimentary facies analysis suggests that the Hualapai Limestone is a spring-fed lake and marsh system sourced by springs located along faults within the depositional basins, analogous to modern Grand Canyon springs and Colorado Plateau groundwaters (not Colorado River water). Stratigraphic units below the lowest carbonates are fanglomerate and evaporites, with different subcarbonate stratigraphy in each basin suggesting initially separate depocenters. The volume of lacustrine carbonate relative to marsh facies carbonates and siltstone increases up-section in each basin, suggesting deepening through time. Sustained groundwater through-flow in the basins of the lake system from 13 – 6 Ma is necessitated by thick carbonates and relative lack of evaporite.

Carbon (C) and oxygen (O) isotopic analyses show up-section changes (fig. 1) over hundreds of meters of section (several million years), from restricted systems with endogenic-dominated spring waters (δ18O=-3.44‰ and δ13C=4.25‰) to open systems with increasing freshwater components and higher elevation recharge (δ18O=-13.00‰ and δ13C=-1.75‰). Heavy C values (0 to +4) overlap with modern Havasu Canyon travertines (O'Brien and others, 2006) and reflect high organic productivity in the lake/marsh system. Radiogenic 87Sr/86Sr values of (0.7114 – 0 0.7193) decrease up-section and are similar to mixtures of springs and groundwaters seen in western Grand Canyon travertines (Crossey and others, 2006, 2009), confirming a freshwater origin for the Hualapai Limestone. We conclude that waters that fed Lake Hualapai were similar to modern CO_2 springs and groundwaters (for example, Havasu Creek) that are fed by mixed, endogenic- influenced groundwaters, rather than Colorado River water.

Figure 2*A*-*C* shows analyses that were performed on detrital zircons from 3 samples in Grand Wash trough. A sample of the siltstone facies that underlies and interfingers with the lacustrine carbonate facies was collected from just above a 13 Ma tuff in the uppermost Muddy Creek Formation in Grand Wash trough: It shows a bimodal age probability plot with modes of 1,740-1,680 Ma and 1,387 Ma (fig. 2A). Samples of fanglomerate of the Muddy Creek Formation (fig. 2B) and red siltstone (fig. 2C) that underlie the ~6 Ma upper Hualapai Limestone of western Grand Wash trough both show similar bimodal age probability plots with modes of 1,740 – 1,650 Ma and 1,380 Ma. All these samples lack the rich spectrum of zircon grains seen in the Phanerozoic strata of the Colorado Plateau (fig. 2F). The relative absence of Phanerozoic grains precludes derivation of the 12 – 6 Ma Hualapai Limestone of Grand Wash trough from the east through postulated paleodrainages draining the Colorado Plateau (Hill and others, 2008; Young, 2008). The 1,387 Ma peak is surprising given the prevalence of 1,450 Ma Gold Butte granite clasts in the fanglomerates underlying the Hualapai Limestone and suggests that Hualapai red siltstone detritus was derived from southern paleodrainages from the Kingman Arch, where 1,380 Ma granitoids are found (Reynolds, 1988), rather than from the northwest, where Gold Butte granite is found. Thus, the detrital-zircon data reinforce models for an internally-drained Grand Wash trough from 13 – 6 Ma, and agree with the "Muddy Creek constraint" (see Lucchitta, this volume) for the lack of any Colorado River/Colorado Plateau detritus from 13 – 6 Ma.

Red siltstones in these western locations contain detrital-zircon age-probability plots (figs. 2*D*, *E*) that are markedly different from time equivalent units in Grand Wash trough reinforcing a model for separate development of each basin. These rocks include zircon peaks of 1,200 – 1,000 Ma and 650 – 550 Ma that are interpreted to have been derived from western provenances. Neither the eastern or western siltstones that just underlie the 6 Ma top of the Hualapai Limestone resemble the Colorado River reference detrital-zircon distribution (fig. 2*G*) reinforcing models for the arrival of the Colorado River after 6 Ma.

Tephrochronologic analyses yields an additional basal date of 12.07 – 11.31 Ma for the Hualapai Limestone of Grand Wash trough. This sample is 11 m above the base of the carbonate section and dates the transition from red siltstone to Hualapai Limestone in the depocenter of the Grand Wash trough as ~12 to ~11 Ma. In other locations, Muddy Creek Formation grades upwards into red siltstone facies, then Hualapai Limestone, time transgressively over the interval ~13 to ~6

Ma. Several tephra were collected near the top of the section in Temple Basin. A sample 24 m below the 5.97 Ma volcanic ash of Spencer and others (1998) correlates with a sample in Detrital Basin, helping to support the model for time-correlative tops of Hualapai Limestone in different basins.

The geochemistry and sedimentology of the Hualapai Limestone provide important constraints for Colorado River integration models. (1) Large volumes of carbonate imply through-flow of water (dominantly nonevaporative conditions) over millions of years. (2) Geochemical signatures similar to modern Colorado Plateau groundwater rules out westward flow of Colorado River water through sink holes and karst (Hunt, 1956; Hill and others, 2008). 3) Gradual up-section chemostratigraphic changes do not support Lake Bidahochi spillover (Meek and Douglass, 2001). 4) Detrital zircon studies rule out: significant western Colorado Plateau paleodrainages (Hill and others, 2008; Young, 2008) and models of headward erosion into the Colorado Plateau from 13 – 6 Ma (McKee and others, 1967; Lucchitta, 2003); and they do not support the existence of pre-6 Ma west-flowing paleodrainages from the Colorado Plateau (Polyak and others, 2008).

We propose a groundwater-sapping model (Pederson, 2001) for Colorado River integration that envisions groundwater-fed springs that helped focus surface drainage to help carve canyons that breeched drainage divides. Water balance and observed hydrochemical changes in the Hualapai Limestone reflected responses to continued 13 – 6 Ma normal faulting and a growing San Francisco volcanic field recharge area. Following initially separate basin evolution, we support models that postulate that waters in the Grand Wash trough became integrated into Gregg Basin, then Temple Basin, then Detrital Basin, to cause integration of the Colorado River ~5 – 6 Ma.

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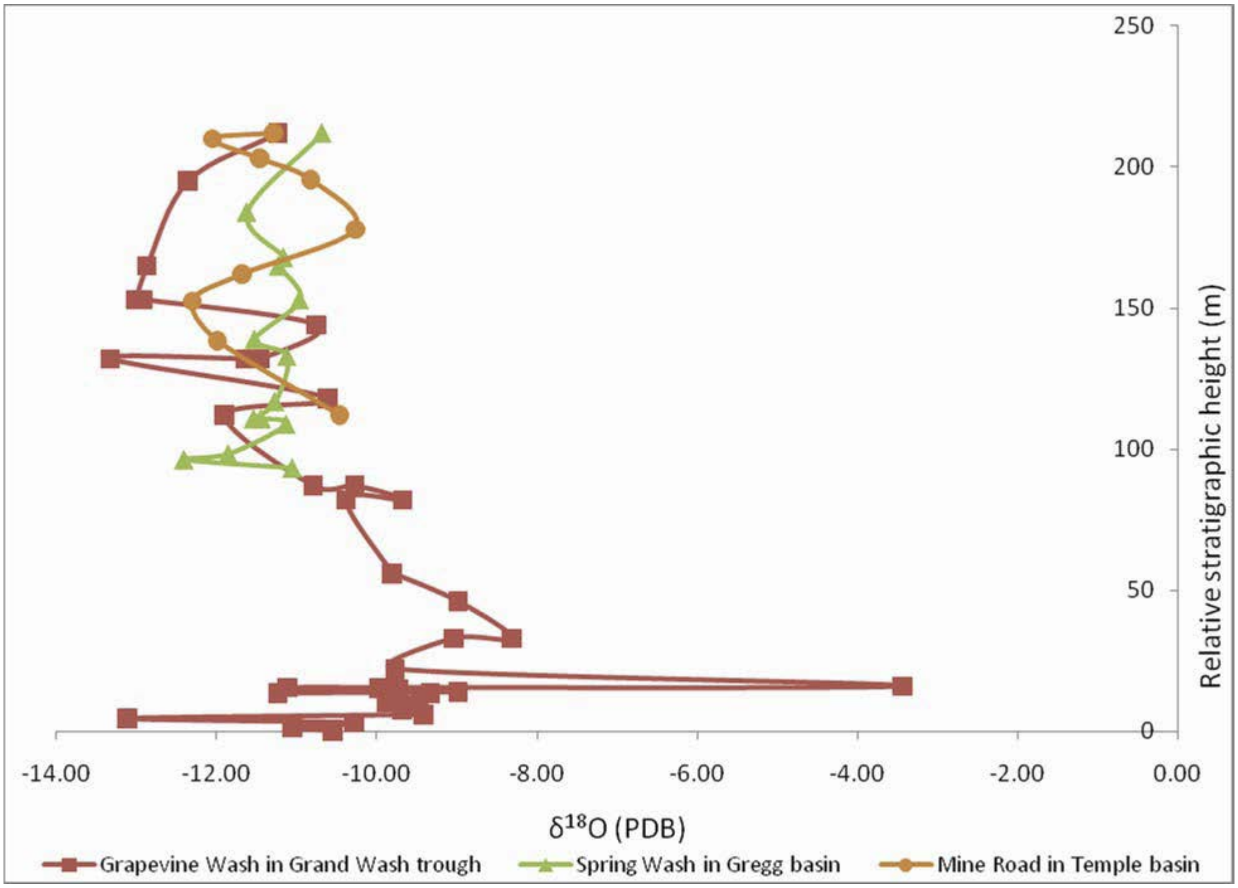

Figure 1. O isotope geochemistry of the Hualapai Limestone. O isotopes show consistent results in different basins and exhibit up-section decrease in δ18O suggesting increased meteoric water input during the several million year (12-6 Ma) duration of deposition of the Hualapai Limestone. Similar stratigraphic trends in carbon isotopes and 87Sr/86Sr values suggest gradual freshening of lake waters.

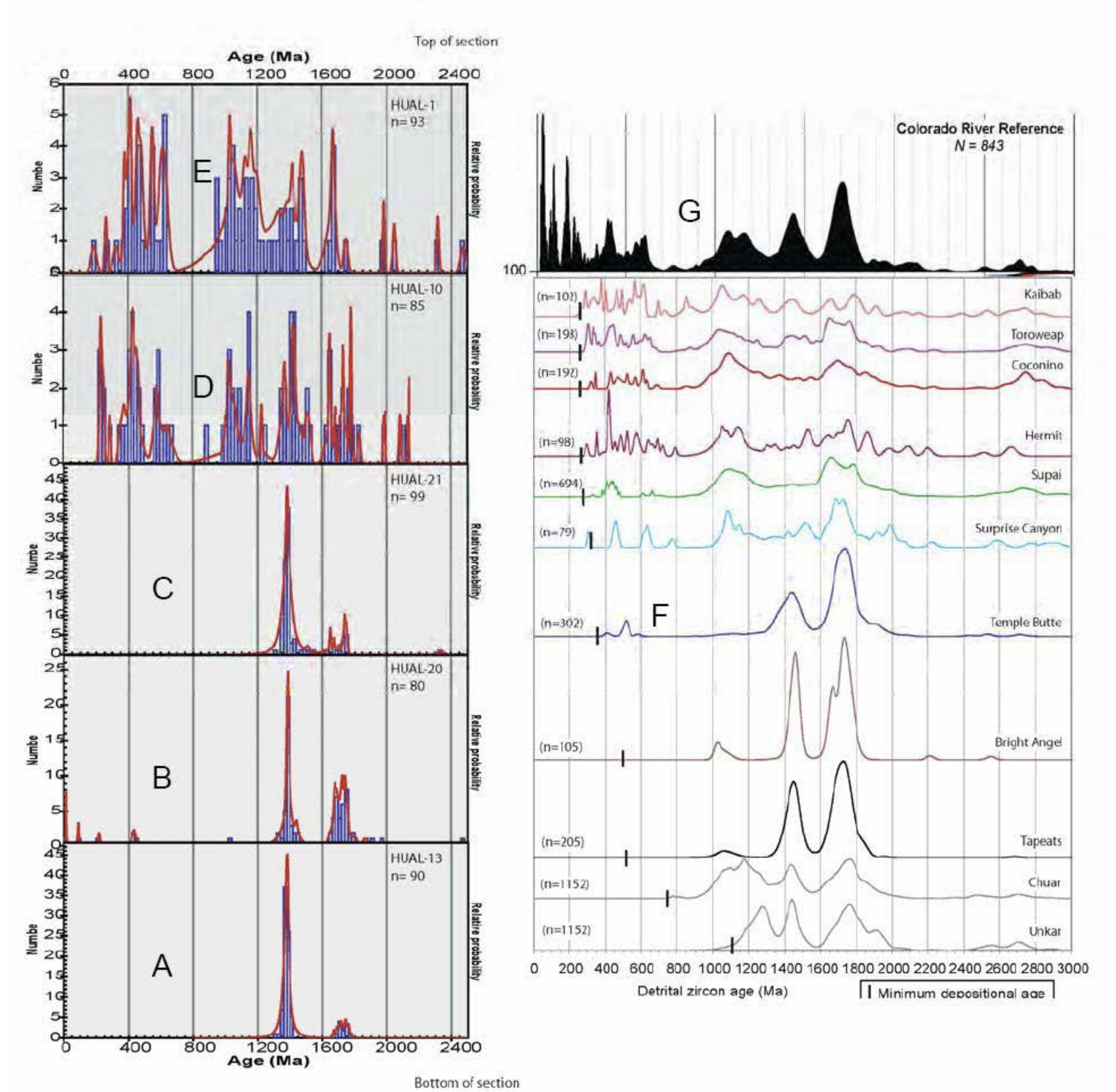

Figure 2. Detrital zircon data from the Hualapai Limestone (A-E) compared to the Grand Canyon Paleozoic reference column (F, from Gehrels and others, in preparation) and the modern Colorado reference distribution (G, from Kimbrough and Grove, in preparation). A) sandstone from the Muddy Creek Formation just beneath the 13 Ma ash near Pearce Ferry; B) Red siltstone from Hualapai Limestone just beneath ca. 6 Ma Hualapai Limestone near South Cove; C) fanglomerate from just beneath ca. 6 Ma Hualapai Limestone near South Cove; D) Red siltstone of lower Hualapai Limestone from Gregg basin; E) Red siltstone of uppermost Hualapai Limestone from Temple basin; F) Reference detrital zircon populations from Paleozoic sedimentary rocks of the Grand Canyon (from Gehrels and others, in preparation); G) Reference detrital zircon population from the modern Colorado River (Kimbrough and Grove, in preparation).

The Miocene(?) Crooked Ridge River in Northern Arizona and its Implications for the Colorado River and Grand Canyon

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The sinuous Crooked Ridge on the Kaibito Plateau of northern Arizona extends continuously from the eastern edge of White Mesa southwestward to The Gap, a wind gap carved into the Navajo Sandstone at the Echo Cliffs (figs. 1 and 2). The ridge is 55 km long along its trace.

Crooked Ridge now stands in inverted relief, which was created because material in and under the channel of an ancient river has been protected by river gravel reinforced by a massive 1–2 m layer of calcrete. In contrast, the rest of the valley was not so protected and has been lowered by erosion. Only remnants of the gravel-cap are preserved, they and extend for about a third of the ridge's length eastward from The Gap and about a quarter of its length westward from White Mesa. The intervening part has either fragmentary gravel caps or no caps at all. The gravel-capped parts of the ridge rise as much as 110 m above the adjacent landscape, whereas the bedrock parts are 50–80 m above it.

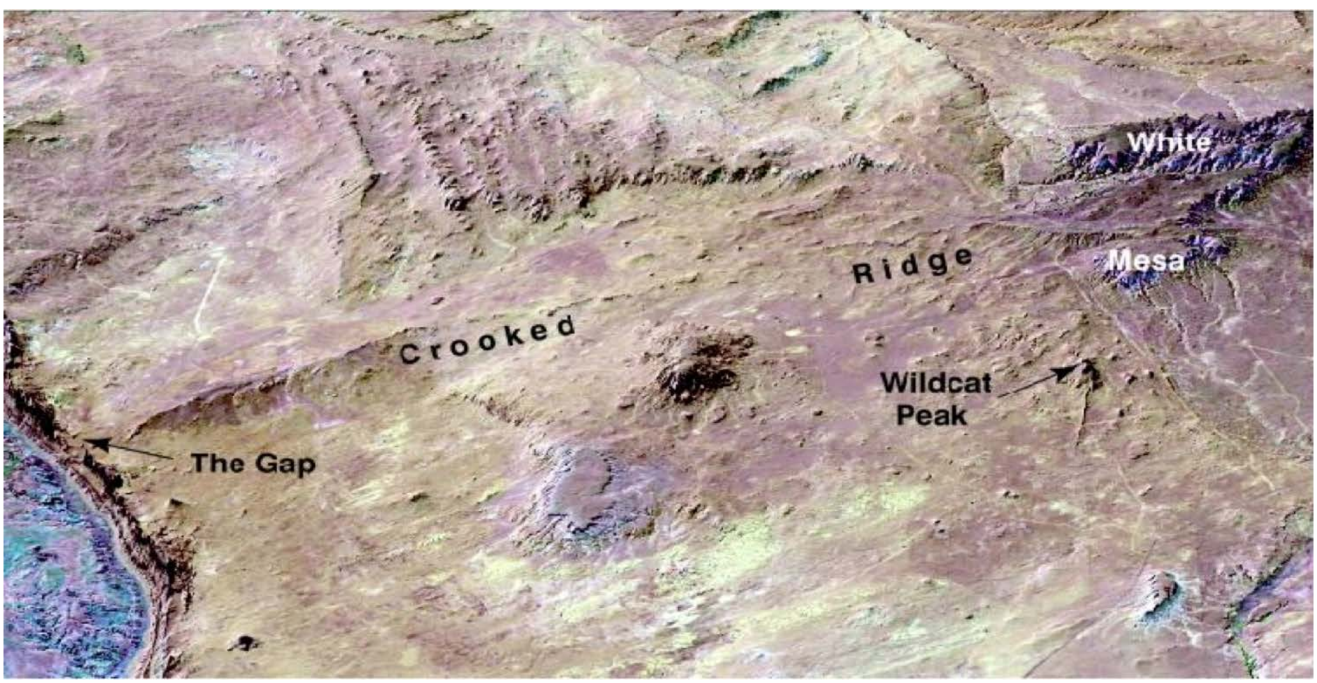

Figure 1. Oblique view, looking north, of Kaibito Plateau and Crooked Ridge. The broad, ancient valley of the river is visible (Composite of Landsat and 10-m shaded-relief DEM).

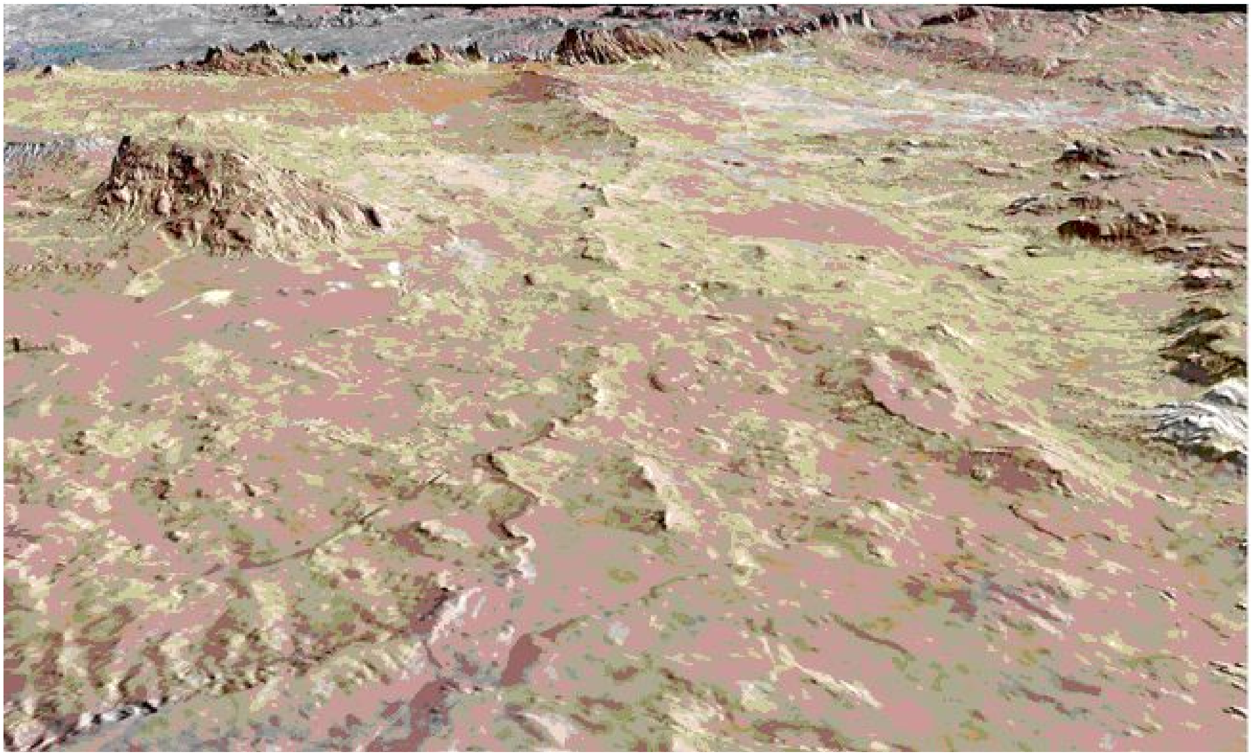

Figure 2. Oblique view looking SW along Crooked Ridge to The Gap, 48 km away in a straight line. The ridge follows the thalweg of the ancient river in inverted relief. Wider parts of the ridge in the left foreground and at far end are mantled with gravel; parts in between are primarily bedrock. (Composite of Landsat and 10-m shaded-relief DEM, 2x vertical exaggeration).

The Crooked Ridge gravel has long been known (Cooley, 1960; Hunt, 1969; Stokes, 1973, Hereford, unpublished data), but a true understanding of the extent, connection, and valley characteristics of the Crooked Ridge drainage became possible only with the advent of 7.5' topographic maps and, especially, high-quality satellite images, which showed the shape of the ridge/channel clearly (fig. 2) and triggered the present study.

Hunt (1969) used the name "Kaibito Plateau gravels", based on information from scattered observations by Cooley (1960). We prefer the name "Crooked Ridge gravels" because of the specific association with the ridge on which they occur.

The farthest-west and lowest gravel is near The Gap at an elevation of about 1,700 m. The farthest east gravel is near the north corner of Black Mesa at an elevation of 2,230 m. The Black Mesa gravel has a similar clast composition as the deposits on Crooked Ridge—especially the exotic far-traveled clasts of quartz metaconglomerates and rhyolitic vitrophyre (see below)—and is on the same map and gradient trend as the rest of Crooked Ridge, so it is unlikely that it was deposited by some other unrelated river.

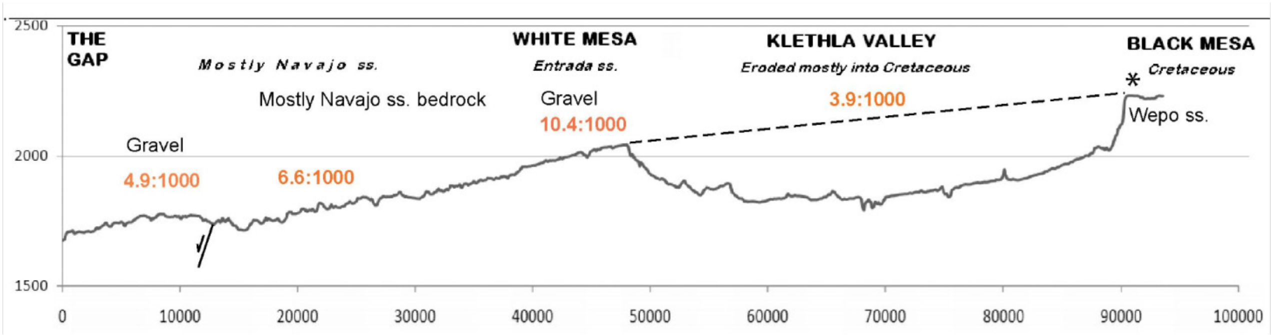

Figure 3. Longitudinal profile along the Crooked Ridge drainage. Horizontal scale is distance along the ridge from The Gap, in meters. Vertical scale is elevation, in meters. Italics indicate bedrock in which the old valley is cut. Roman lettering below indicates material on ridge. Numbers in red indicate the gradient of the river (m/1000m). * Indicates location of gravel deposit on Black Mesa. (DEM data, 30-m sampling interval).

The total length over which Crooked Ridge River (CRR) can be traced from Black Mesa to The Gap is 91 km. Over this distance, the river drops 530 m, giving it an average gradient of 5.8 m/km. The gradient of individual reaches varies considerably, probably reflecting the rocks into which the valley was carved, structural features, such as monoclines and faults, and constrictions in the river's path, such as The Gap (fig. 3). The high gradient and the coarsening-upward sequence of the deposits can be attributed to overloading with sand, silt, and clay derived from the easily-eroded Cretaceous rocks that formed the valley sides on and upstream from the Kaibito Plateau.

An alternative explanation for the high gradient is post-depositional tilting of the channel owing to crustal-warping mechanisms, such as isostatic unloading to the north (Lazear and others, this volume) and mantle plumes (Robert and others, this volume; Moucha and others, 2009). However, the proposed models are not in close agreement and do not take into account the progressive post-Miocene south-to-north stripping of Cretaceous strata on the southern Colorado Plateau (Fleming, 1994). The models also do not account for the erosional retreat of scarps northeast down the structure at a rate of 4–8 km/Ma (Lucchitta, 1984, 1989; Holm, 2001; Lucchitta and Jeanne, 2001), which would cause removal of load and, therefore, isostatic rebound in the south earlier than in the north. Finally, the course of CRR is at a low angle to the contours of isostatic-rock uplift shown by Lazear and others (this volume), so the post-depositional tilting due to rebound would be small.

Hereford (written commun., 2010) points out that the gradient of the modern San Juan River is much less than that of CRR, suggesting tilting. However, the modern San Juan River is a mature drainage, whereas CRR probably was not, considering the overloading and disturbances that took place upstream from Black Mesa in Crooked Ridge time and shortly before. These include emplacement of the Navajo volcanoes (28–19 Ma) and of some of the laccolithic intrusions (29–20 Ma), as well as possibly late phases of the San Juan Mountains volcanism (35–15 Ma).

We believe that the steep gradient of CRR is better explained by overloading than by post-depositional tilting, although minor tectonic adjustments cannot be ruled out.

The sinuosity ratio of CRR is 1.15. The meander amplitude is 110–380 m, the floodplain width is 410–850 m, and the inferred width of the valley is possibly as much as 8.5 km. The valley has been widened by erosion, but it is still conspicuous on the images, suggesting minor modification. The rim-to-rim width of the wind gap in the Echo Cliffs at The Gap is 3 km.

The parameters of CRR compare well with those of rivers in the region with considerable discharge (Animas River upstream from Durango, Colorado, mean maximum daily discharge of 85 m^3/s and a maximum daily flood over 99 years of record of 303 m^3/s; and San Juan River upstream from Pagosa Springs, Colorado, mean maximum daily discharge of 45 m^3/s and a maximum recorded daily flood of 130.7 m^3/s). This supports the interpretation that CRR was a substantial river, not a minor one.

Composition, Source, and Rounding of Clasts

Table 1 lists the composition of clasts collected. The inferred sources of the clasts collected from the gravel deposits in various localities on Kaibito Plateau and from Black Mesa are shown in table 2. Rounding of clasts ranges from angular (very few) to well rounded. Most locally derived clasts are subangular to subrounded; most exotic clasts are subangular to well rounded.

Table 1. Composition of Gravel Samples

Statistical Samples

Lithology	Percent CR2-2[1]	Percent CR2-3[2]
Mesozoic sedimentary[1]	53	61
Quartzite	13	19
Chert	12	5
Felsic volcanic[2]	5	7
Minette	9	2
Granite	5	1
Intermediate volcanic[3]	1	2
Quartz	1	1
Metaconglomerate[4]	0	1
Gneiss[5]	0	1
Earthy hematite	1	0

[1]Sandstone, conglomerate, siltstone, and claystone.
[2]Rhyolite, argillic-altered rhyolite, vitrophyre, welded tuff, and compacted tuff.
[3]Andesite and latite.
[4]Pebbles are quartz and quartzite.
[5]Fine grained, linear, and quartzofeldspathic.

Nonstatistical Samples

Lithology	Percent[3]
Quartzite	48.5
Chert	11.5
Mesozoic sandstone[1]	7.3
Metaconglomerate[2]	5.2
Quartz	5.2
Granite	4.6
Metawacke	4.0
Microcline	3.4
Intermediate porphyry[3]	2.9
Felsic volcanic[4]	2.9
Minette	2.3
Intermediate volcanic[5]	1.1
Earthy hematite[6]	1.1

[1]Only fossiliferous and iron-manganese cemented.
[2]Pebbles are quartz and quartzite.
[3]Hornblende and plagioclase phenocrysts; andesitic to dacitic (dioritic to granodioritic).
[4]Rhyolite and vitrophyre.
[5]Andesite and propylitic-altered andesite.
[6]With calcite veins.

[1]CR2-2: 100 contiguous pebbles from the top surface of the deposit at big quarry, Kaibito road, White Mesa.
[2]CR2-3: 100 contiguous pebbles from the bottom of the vertical face in big quarry, Kaibito road, White Mesa.
[3]Percent of 175 pebbles and cobbles picked up randomly and individually (cherry picked) at seven sites, includes 68 specimens from Black Mesa. The Black Mesa specimens include quartzite, quartz, chert, granite, pegmatite, rhyolite, rhyolitic vitrophyre, and quartz metaconglomerate; the last two are Distinctive Lithologies in table 2.

Table 2. Inferred Sources of Pebbles and Cobbles

[●, Probably; x, Possibly; m, Maybe; (375), number of specimens collected. NM, Needle Mountains; SJVF, San Juan volcanic field; LACC, Colorado Plateau laccoliths; 4CDD, Four Corners dikes and diatremes; CH, Chinle Formation; MO, Morrison Formation; J–K, Jurassic and Cretaceous strata.]

Distinctive Lithologies	NM	SJVF	LACC	4CDD	CH	MO	J-K
Compacted rhyolitic crystal tuff (1)		●					
Argillic-altered rhyolite (3)		●					
Rhyolitic vitrophyre (2)		●					
Latite (2)		●					
Andesite (2)		●					
Propylitic-altered andesite (1)		●					
Earthy hematite with calcite veins (3)		●					
Quartz metaconglomerate (10)	●				m	m	
Metawacke (7)	●						
Fine quartzofeldspathic gneiss (1)	●			m			
Hornblende-plagioclase porphyry (5)			●				
Minette (15)				●			
Sandstone (gray, tan, brown) (108)							●
Conglomerate, siltstone, claystone (7)							●
Tan sandstone with shells (12)							●
Petrified wood						●	
Nondistinctive Lithologies							
Quartz (11)	x	x		x	m	m	
Quartzite (tan, gray, red, yellow) (117)	●				m	m	
Chert (37)	m				m	m	m
Rhyolite (8)		●			m	m	
Felsic welded tuff (3)		●			m	m	
Granite and pegmatite (red, tan) (13)	x			x		m	
Fine hypidiomorphic-granular granite (gray) (1)	m	m	x			m	
Microcline (red, perthitic) (6)	x			x		m	

Taken together, the lithologies clearly indicate a source to the northeast, and perhaps north-northeast, at least as far as the San Juan Mountain region of Colorado (also see Cooley, 1960, Hunt, 1969, and Stokes, 1973). Metamorphic rocks, like those of Crooked Ridge, currently are exposed in the Needle Mountains region of Colorado at altitudes of 4,000 m or more (Barker, 1969). The hornblende-plagioclase porphyry clasts are derived from the several shallow laccolithic intrusions in the Four Corners region (Emery, 1916; Ekren and Houser, 1965; Witkind, 1964), but their relative scarcity contrasts markedly with their abundance in even the oldest terraces of the Colorado River. These observations suggest that, when CRR was active, the intrusives were exposed much less than they are today and some not at all (Eckel and others, 1949). Today, they are exposed at altitudes as high as 4,000 m, and many are at 3,000–3,500 m, so the ancient topographic surface must have been even higher. In CRR time, it is likely that only the highest were exposed. Today, much of the region in southwest Colorado and southeast Utah near the intrusives is at 1,500–2,000 m elevation, indicating removal of 1–2 km of strata.

The gradient of CRR gives another means for estimating the erosional lowering. The straight-line distance from the Black Mesa gravel outcrop to the Needle Mountains is 300 km. Over this distance, the river bed rose nearly 1,700 m above the 2,230 m altitude of the Black Mesa deposit, if one uses the average gradient (in reality, the gradient may have steepened headward as is typical of river profiles). The resulting 3,900 m altitude at the Needles Mountains allows tapping the Proterozoic basement, but not most of the laccolithic intrusives.

Time of Deposition of Clasts

We have no direct way of dating Crooked Ridge. One indirect method is based on the observation that the course of CRR has no relation to the present drainage network, and that two distinct episodes of drainage development have occurred since Crooked Ridge time. The first episode is the development of Klethla Valley (fig. 3), which truncates and bisects the course of the ancient river. Klethla Valley is mature and typical of many such valleys on the Colorado Plateau. Klethla Valley, in turn, is being beheaded by immature drainages of the canyon-cutting cycle that are tributary to the Colorado and San Juan Rivers (figs. 3 and 4). Thus, the two major rearrangements of the drainage network in the area since Crooked Ridge time suggest an indeterminate but major time interval. Similarly, CRR was not incised, whereas Pliocene and younger streams of the region are. This indicates a pre-Pliocene age.

Another indirect method involves Wildcat Peak (fig. 1), a monchiquite intrusive that is part of the Tuba volcanic field (Akers and others, 1971). The eroded top is now at ~2,030 m. Presumably, there was originally a volcanic edifice above the neck, as is typically inferred for necks and diatremes in the Four Corners region. Drainage from Wildcat Peak to the west-flowing CRR would have been to the north or northwest. CRR is 12–13 km away in this direction and currently at 1,900–1,950 m, so drainage from Wildcat Peak to CRR was possible and likely. Nevertheless, no monchiquite clasts have been found in the gravel. Wildcat Peak has not been dated, but volcanic rocks of this composition have been dated in the Hopi Buttes at 6–8.5 Ma (Damon and Spencer, 2001). Therefore, we infer that the river ceased to function before the intrusive was emplaced in the late Miocene

A third method involves the removal of 1–2 km of strata needed to unroof the laccolithic intrusions northeast of Black Mesa in post-Crooked Ridge time. Such deep erosion is in agreement with results obtained by different techniques and likely involved a substantial time interval.

On the basis of these considerations, we infer that CRR was active in mid-Miocene (and perhaps earlier?) time, in keeping with Hunt's (1969) proposal, and became inactive in pre-Pliocene time. Future dating of the calcium carbonate in the caliche cap may provide a minimum age for the time when the river became inactive.

Regional Drainage Implications

Given the southwesterly course of CRR from the San Juan Mountains to the Kaibab Plateau, no other river could have flowed from the north across CRR to empty into Hopi Lake, nor could Hopi Lake have drained northward across CRR.

The river gravel is at 1,700 m near The Gap. This elevation gives us a means for identifying the terrain across which CRR could not have continued downstream, assuming no major tectonic adjustments (fig. 4). Specifically, the Hopi Buttes and Bidahochi basin are part of the excluded terrain because the base of the Bidahochi Formation as exposed is at ~1,750 m (Love, 1989;

Cather and others, 2008), so CRR could not have filled the hypothetical Hopi Lake, even assuming no gradient to the river. The straight-line distance from The Gap to the nearest side of the Hopi Buttes is ~130 km, giving a drop of 750 m using the average gradient. This would place the river at 950 m in the area. Even half the gradient would make the river much too low to account for the basal Bidahochi Formation and even less likely to account for the higher parts of the section.

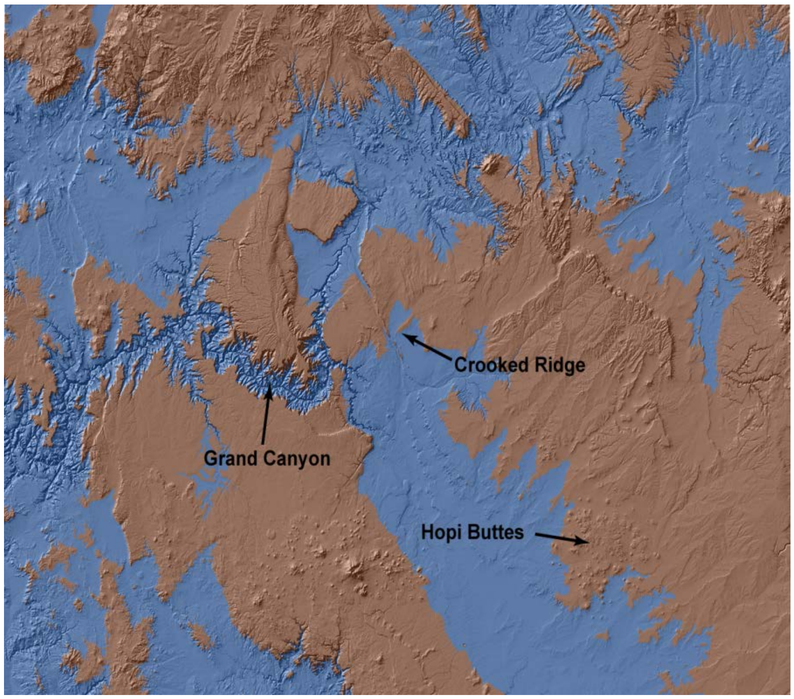

Figure 4. Terrain above 1700 m (brown) and below (blue). DEM data

Figure 4 also shows that the only possible continuation from The Gap would have been southward along a strike valley parallel to the present Echo Cliffs, then along the alignment of the present gorge of the Little Colorado River to near the present confluence with the Colorado River. From there, CRR could continue only north along the alignment of present Marble Canyon, or west along the alignment of present eastern Grand Canyon. If the ancient river flowed north, the distance from The Gap to Lees Ferry along this course is 135 km. Applying the 5.8 m/km gradient of CRR gives a drop of about 775 m over the distance, for an elevation of 925 m, which is below even the present ground elevation at Lees Ferry (the Colorado River is at ~944 m just below the Paria River riffle). Furthermore, the current elevation along the Colorado River is the result of much erosion since Crooked Ridge time. We know that, at Johnson Point just north of Lees Ferry, the bedrock strath beneath an old gravel terrace of the Colorado River is at 1,135 m. This gravel contains a Stage V carbonate soil, which we interpret as 525–600 ka on the basis of correlation with a well-dated Stage V carbonate at River Mile 207.5 (Lucchitta and others, 2000). Assuming a 9 m depth to bedrock in the modern river channel, the strath-to-strath lowering has been ~200 m in the past 525–600 ka (Lucchitta, unpublished field data; Lucchitta and others, 2001), giving an incision rate of 380–330 m/Ma. Even if the rate of incision has not been constant over longer periods of time, the topographic elevation in the Lees Ferry area would have been much too high to be a continuation for a mid- to late-Miocene CRR. On the other hand, our data do not preclude the possibility of a river (ancestral Colorado?) flowing southward along approximately its present course, but at a higher elevation, and joining the CRR near the present-day confluence of the Little Colorado River.

A course westward is possible along the present alignment of eastern Grand Canyon. This potential route was recognized long ago by Babenroth and Strahler (1945) and Lucchitta (1975, 1984, 1989), who viewed it as a valley that follows the curving strike around the nose of the south-plunging North Kaibab upwarp. The greater width and complexity of the Grand Canyon here also corroborates an older age than that of other parts of the Canyon. Most likely, the old course was in a broad valley incised some hundreds of meters below the present Kaibab Limestone rims. An old age for the eastern Grand Canyon is supported on different grounds by Flowers and others (2008).

Once across the Kaibab upwarp, the river(s) might continue along a northwest-trending strike valley, as proposed by Lucchitta (1975, 1984). The analysis of the further course is not within the purview of this paper.

Conclusions

1. CRR was a major stream of the region that can be traced from The Gap northeast at least as far as the San Juan Mountains, and possibly farther north. The age of the river is poorly constrained. In keeping with Hunt (1969) and our findings, we think it was probably of Middle Miocene age, and possibly older. The river probably became inactive in pre-Pliocene time.
2. After the river became inactive, the Four Corners region was lowered erosionally by many hundreds to a few thousand meters.
3. The ancient river bears no relation to the present-day drainage network.
4. The river could not flow into Hopi Lake or northward along the present course of Marble Canyon.
5. Another river could not flow southward across CRR into Hopi Lake, nor could Hopi Lake drain northward across CRR.
6. A river (ancestral Colorado?) could flow southward along the present alignment in Marble Canyon to join CRR near the present Little Colorado River confluence and then continue across the Kaibab Plateau.
7. A westward course across the Kaibab Plateau along the present eastern Grand Canyon is possible, and it is our favored alternative.

Acknowledgements

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Dedicated to
Charlie Hunt, Chester Longwell, and Eddie McKee
The pillars on which we stand

The Muddy Creek Formation at the Mouth of the Grand Canyon—Constraint or Chimera?

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From the time of J.W. Powell (1875), and especially Dutton (1882), until the early years of the twentieth century, geologists viewed the Colorado River as having had the same course since its birth, which was placed early in Tertiary time.

In the 1930s, work in the Basin and Range Province along the lower Colorado River corridor revealed many interior-basin deposits athwart the course of the river. These deposits are Miocene, so no Colorado River could have flowed there until the end of interior-basin deposition, which is generally placed near the end of Miocene time. This conclusion was especially true of Longwell's (1936) work in the future Lake Mead area, notably near Pierce Ferry at the very mouth of Grand Canyon (fig. 1), where the Colorado River has cut about 600 m into the Muddy Creek Formation, exposing details of the stratigraphy and facies relations. According to Longwell, both provide evidence against a through-flowing river. Thus, there seemed to be an ancient upper Colorado River system with no continuation west of the Colorado Plateau.

In the middle of the century, the geologist C.B. Hunt evaluated and compiled all the available evidence for ancient river systems in western Colorado, eastern Utah, and northern Arizona. Hunt (1969) concluded that the evidence supported the notion of ancient river systems, such as an ancestral San Juan, that did not depart much in overall arrangement from that of the present rivers. He also suspected that the ancient rivers were active in middle Miocene or earlier time, although the evidence available to him was not conclusive.

Hunt suggested that this river system flowed across the south end of the Kaibab upwarp and then westward, generally along the present Grand Canyon as far as Peach Springs Canyon. He postulated that the old river then flowed southward, rather than northward, as the stream in Peach Springs Canyon does today, thus conveniently bypassing the Pierce Ferry area. To establish the present course of the Colorado River in western Grand Canyon and the Basin and Range reach, Hunt proposed that the river began discharging into the Pierce Ferry area by means of subterranean piping at the end of Muddy Creek time. This created a lake in which was deposited the Hualapai Limestone—the uppermost unit of the Muddy Creek Formation.

E.D. McKee was skeptical about Hunt's proposal. He therefore persuaded two graduate students to examine two critical areas in detail: Peach Springs Canyon and the Pierce Ferry area.

R.A.Young studied Peach Springs Canyon and the Hualapai Plateau. Using clast provenance, as well as the internal structures and morphology of the gravel deposits, he showed that drainage in the canyon has been to the north since early Tertiary time, thus blocking Hunt's convenient escape route (Young, 1979, 1982).

I. Lucchitta (1966, 1967, and 1972) mapped the Pierce Ferry area together with much of the Grand Wash trough and confirmed Longwell's earlier, less detailed work. There was no evidence for a Grand Canyon during or before Muddy Creek time.

Fortified with these results, McKee convened a symposium in 1964. After deliberating, the participants (McKee and others, 1967) accepted both the antiquity of the upper drainage and the obstacle at the mouth of the Canyon. Since the Kaibab upwarp seemed a formidable impediment, it was proposed that the old stream flowed along the present Little Colorado River valley, but in the opposite (southeast) direction, joining the Rio Grande on its way to the Gulf of Mexico. Meanwhile, the Hualapai river system could exit southward through Peach Springs Canyon as suggested by Hunt (Young had not yet finalized his work). After the opening of the Gulf of California 5–5.5 Ma, a youthful stream eroded its way headward from the Gulf into the Colorado Plateau along the present Grand Canyon (following strike valleys and structural features), capturing the Hualapai River and then the old upper river east of the Kaibab Plateau. The modern river was born and cutting of the Canyon began.

However, there is no evidence for a river flowing southeast along the present alignment of the Little Colorado. Lucchitta (1975, 1984) then suggested that the ancient river could have crossed the Kaibab upwarp in an arcuate strike valley coincident with the present eastern Grand Canyon, but shallower, as proposed earlier by Babenroth and Strahler (1945). The river would then continue west of the Kaibab upwarp in a northwest-trending strike valley, later to be captured west of the upwarp by the new lower river draining into the Gulf of California.

Both mechanisms depend heavily on headward erosion. These ideas represented a major change in thinking because drainages were seen not as fixed, but as part of constantly evolving networks whose connections change in response to external factors such as tectonism, and principally through headward erosion and stream capture.

The last quarter century has seen numerous theories and ideas regarding the history of the Colorado River and its integration into the present course (for example, Robert and others, this volume; Wernicke, 2011; Faulds and others, 2001; Wallace and others, 2005; Young, 2008; Wernicke, 2011). Subterranean piping is a popular theory (Hill and others, 2008; Pederson, 2008), as is lake spillover (Blackwelder, 1934, Meek and Douglass, 2001; Scarborough, 2001; Spencer and Pearthree, 2001). Another idea suggests that parts of the Canyon are old and were occupied by the Colorado River, but were choked by debris during interior-basin deposition to the west, so the river became inactive (Elston and Young, 1989), or were formerly occupied by rivers that flowed in directions (generally north) other than the present one (for example, Scarborough, 2001). The name "Muddy Creek" has been abandoned for interior-basin deposits in the Grand Wash Trough, which includes the Pierce Ferry area, on the grounds that no physical connection of these deposits with those in the type locality along Muddy Creek in Nevada can be demonstrated (Bohannon, 1984). However, the change in nomenclature does not change the import of the interior-basin deposits at the mouth of Grand Canyon.

Selected Hypotheses about the Mouth of Grand Canyon

Faulds and others (2001) and Wallace and others (2005) suggest that a westward-flowing stream may have begun to erode headward into the edge of the Colorado Plateau at the site of the present Grand Canyon once faulting had produced a substantial scarp ~13 Ma. This canyon may have been the extension of a paleocanyon carved into the rocks of Wheeler Ridge to the west (fig. 1) before or early in the movement of the Grand Wash fault. As evidence for the canyon, they adduce the red color of the fine-grained sediments near Pierce Ferry, which contrasts with the generally gray tint of the alluvial fans derived from the west. The red color indicates derivation from Pennsylvanian and Permian rocks, found only on top of the Grand Wash Cliffs (or at the north end of Grand Wash trough, not in communication with the Pierce Ferry area). Therefore, the sediments were derived from the east, which is taken as evidence for a Muddy-Creek-age canyon or drainage cutting into the face of the Grand Wash Cliffs.

A variation of this argument, presented by R.A. Young (2008), also holds that a substantial "precursor" canyon was carved at the mouth of the Grand Canyon in Muddy Creek time, after a fault scarp was formed. This canyon would have left no trace in the deposits at Pierce Ferry because it was occupied by a lake that trapped sediment.

Furthermore, the lake would have maintained itself for a long time because the supply of sediment to the lake was low owing to its alleged carbonate-dominated drainage basin and the ephemeral character of many of the tributaries to the paleoriver. A third hypothesis, presented by J. Pederson (2008), is that a paleoriver river discharged by subterranean piping to various springs in the country west of the Colorado Plateau and north of Lake Mead, instead of through a single surface channel.

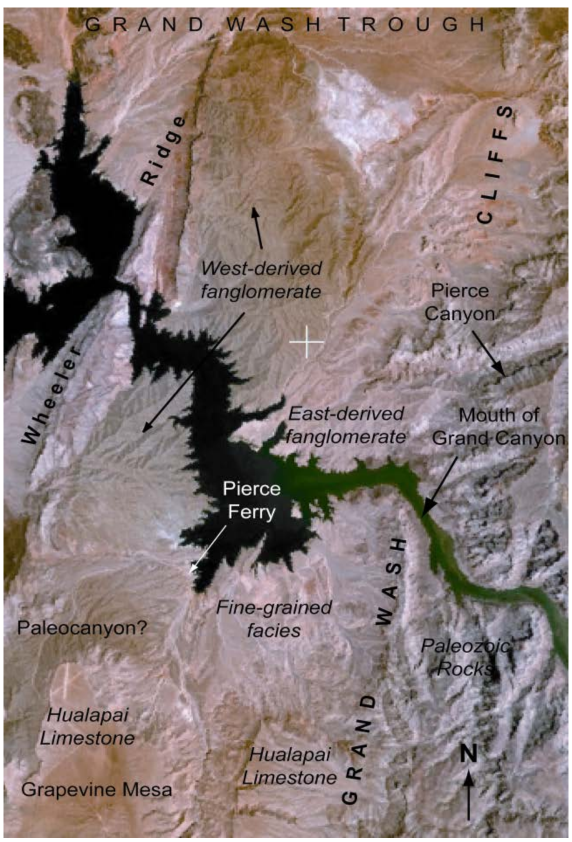

Figure 1. Landsat image showing geographic and geologic features, Grand Canyon area, Arizona.

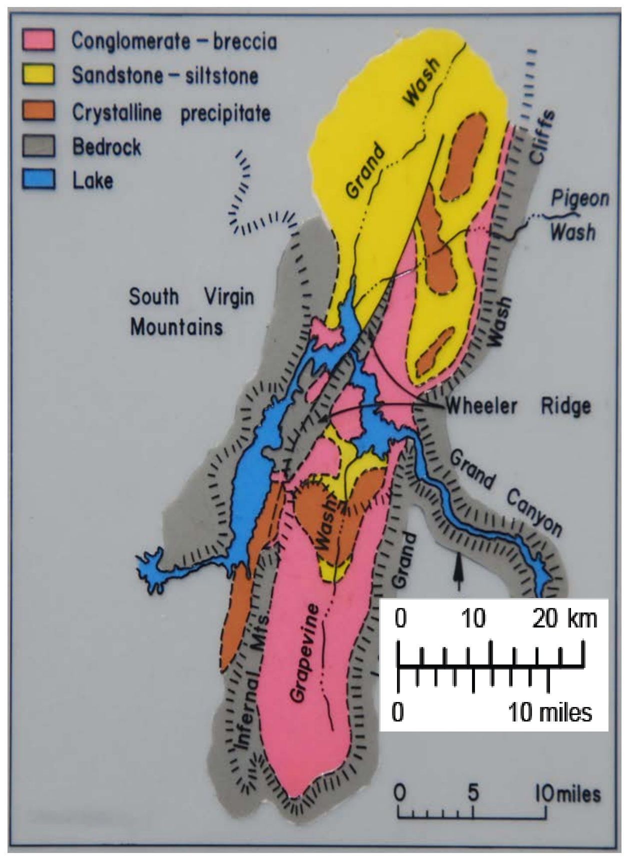

Figure 2. Distribution of facies in the Grand Wash trough near the Pierce Ferry area, Arizona.

Observations

Before movement on the Grand Wash fault, the area southwest of the Colorado Plateau was a topographic and structurally high region. It is the northwest continuation of the Mogollon Highlands. Streams flowed from the uplift onto the Plateau in broad valleys, many of which contain gravel and volcanic rocks (Young, 1979, 1982, 1987; Beard and Faulds, this volume). Among these are the Peach Springs Tuff, 18.4 Ma, and various basalt flows, notably the 17.4 Ma Iron Mountain Basalt, remnants of which are less than 5 km from today's Grand Canyon. All predate movement on the fault. The old valleys are conspicuous on the crest of the northwest-trending Grand Wash Cliffs, but are absent from the northeast-trending part, which includes the mouth of Grand Canyon. Here, the crest is remarkably uniform and broken only by two narrow canyons draining southwest (in addition to the mouth of Grand Canyon itself). The northwest-trending part of the Grand Wash fault started moving after 17.4 Ma, and perhaps the northern segment did the same. Recent work (Faulds and others, 2001) indicates that faulting and tilting occurred primarily 16–13 Ma.

Faulting disrupted the northeast drainage onto the Plateau, formed the scarp of the Grand Wash Cliffs, and created the large Grand Wash-Hualapai Valley tectonic trough at their base. Debris-flows filled the trough with coarse fanglomerate derived largely from the west. This fanglomerate forms conspicuous cones (fig.1) and grades laterally into and interfingers with increasingly fine clastic deposits toward the east. However, Longwell (1936, p.1,434) states that *"Near the river level the granitic material in the fanglomerate is predominant as far east as the old Pierces Ferry, less than a mile from the mouth of the Grand Canyon."*

This observation would place the west-derived fanglomerate in direct contact with the east-derived fanglomerate issuing from Pierce Canyon, leaving no room at the pre-lake river level for fine-grained facies in that area. The finer-grained facies visible today near Pierce Ferry are stratigraphically above the relation described by Longwell.

Material derived from the Grand Wash Cliffs is minor because the scarp is composed of tough carbonate rocks that dip northeast away from the scarp, reducing the amount of water available to erode the scarp and produce valleys (Lucchitta, 1966, 1987). However, the two canyons that are present in the scarp (Pierce and Pigeon Canyons) did produce conspicuous alluvial cones that grade westward into fine-grained basin beds. Pierce Canyon, is directly north of the mouth of the Canyon (figs. 1 and 2).

Mudstone, claystone, and minor impure evaporites and limestone crop out in the lowest, axial part of the old basin near Pierce Ferry. West of the basin axis and on an east-facing paleoslope, fine-grained, reddish-gray material grades into and interfingers with the west-derived fanglomerate.

Both the purity and the areal extent of the limestone increase upwards in the section along the axis of the old basin. The depocenter of the limestone in the Pierce Ferry area is not at the mouth of the Grand Canyon, but south of it (fig. 2). Exposures on the face of Grapevine Mesa just south of Pierce Ferry show that as the basin filled and the influx of clastic material decreased, the limestone transgressed over other rock types (fig. 3). The limestone is mostly impure and contains features, such as plant-stem impressions, suggesting that the lake in which the limestone was deposited was not deep. Neither the limestone nor the other basin beds in the Pierce Ferry area contain evidence of a clastic influx from a paleo-Grand Canyon to the east.

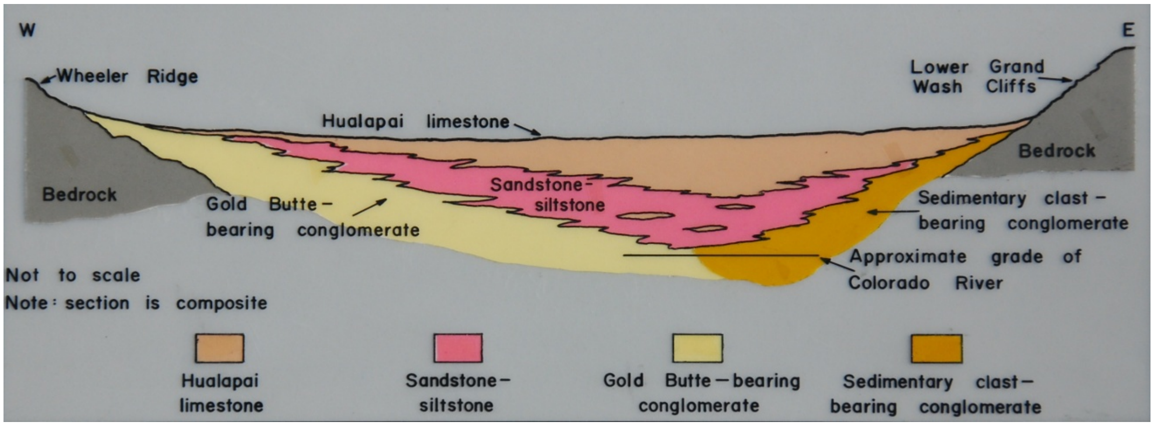

Figure 3. Diagrammatic crosssection of the basin fill near Pierce Ferry, Arizona.

My interpretation of the data has been that the basin initially was filled mostly with west-derived coarse clastic material that overwhelmed everything else. The axis of the basin was near the foot of the Grand Wash Cliffs and was occupied by small playas that expanded as faulting and relief waned and the coarse-grained contribution decreased. Small, shallow ponds or lakes rich in dissolved carbonate and sulfates occupied the lowest areas. The water probably was ground water originating from the plateau to the east and the south Virgin Mountains to the west. Such arrangements can be observed in the present Basin and Range province. As the clastic contribution decreased further, the lake(s) expanded, becoming wetter and cooler toward the end of Miocene time. Eventually, the lakes were tapped and drained by a river working its way upstream from the newly opened Gulf of California. Exposed Muddy Creek sections contain no evidence of sediment contribution or facies distribution that would indicate a Grand Canyon of Muddy Creek age.

This background generates numerous questions that should be addressed by the various hypotheses on how Grand Canyon developed:

1. At the now-obscured river level, there are no fine-grained sediments that could be attributed to a paleocanyon cut into the Grand Wash Cliffs. Did the alleged canyon only come into being later?
2. The fan issuing from Pierce Canyon is reddish and contains material derived from Upper Paleozoic strata exposed north of the present Grand Canyon. The fanglomerate issuing from this canyon was a major supplier of reddish material to the fine-grained sediments in the Pierce Ferry area with which it interfingers. The presence of reddish material does not necessarily imply a paleo-Grand Canyon.
3. The fan issuing from Pierce Canyon is present on both sides of the present-day Colorado River (fig. 4), so no paleocanyon could be there in Pierce Canyon time.

Figure 4. Pierce Canyon fan and the mouth of the Grand Canyon, Arizona.

4. Pierce Canyon left abundant evidence for its presence in Muddy Creek time. Is it reasonable to propose that a paleo-Grand Canyon left none?
5. Why would a paleocanyon of any size develop in a stretch of the Grand Wash Cliffs where otherwise few if any sizable canyons are visible today, and those that are present are visible in Muddy Creek deposits? Why are neither gravel nor volcanic rocks exposed in or near the alleged paleocanyon, as in other paleocanyons on the Hualapai Plateau?
6. If the paleocanyon developed only later in Muddy Creek time, and was occupied by a lake in which the Hualapai Limestone was being deposited, why is no limestone depocenter present at the mouth of the Canyon? If this lake was the sump into which an ancestral upper drainage emptied, where did this drainage go before limestone deposition began in the Grand Wash trough?
7. If a lake was ponded upstream from the mouth of the canyon, what dammed it?
8. Lakes are ephemeral, especially a very narrow one in a canyon. We have learned a lot about the likely lifespan of lakes in the canyon from the Pleistocene lava dams and associated lakes. The proposed lake would have had to last several millions of years (~13–5 Ma) to explain the absence of material derived from the paleocanyon in the section exposed at Pierce Ferry. Is this realistic?
9. The lake is said have been long-lasting because little clastic sediment would have been supplied by the carbonate strata inferred to form the topographic surface over much of the drainage basin at the time. However, even today widespread mid-Miocene to Pleistocene lavas overlie the erodible Moenkopi Formation in the drainage basin near western Grand Canyon, showing that a good supply of sediment from the Moenkopi, and probably from higher Mesozoic units as well, was available to the alleged lake.
10. The notion that tributary streams would have been ephemeral and thus would have contributed little sediment is negated throughout the length of Grand Canyon, where ephemeral streams contribute a great deal of material to the river by means of flash floods. This is the material that forms the rapids of the Colorado River. Base flow is not needed to contribute material.

11. The idea that the ancestral river fed Hualapai Lake by means of springs raises two problems. First, the distribution of the limestone does not indicate point sources, such as springs. Is it not reasonable to think that the water came largely or entirely from groundwater flowing into the lowest parts of a structural and topographic trough? Second, where did the ancestral river go before Hualapai Lake time? Older beds contain little or no limestone. Was there no subterranean water emerging as springs then?

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Observations of the Bouse Formation in Chemehuevi and Parker Valleys, California and Arizona

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Several features in Chemehuevi and Parker valleys near the lower Colorado River are relevant to understanding deposition of the Pliocene Bouse Formation and its bearing on the evolution of the Colorado River in the Basin and Range Province. Central parts of Chemehuevi and Parker valleys expose cross-bedded sand and gravel interfingered with, or immediately beneath, limestone forming the lowest part of the Bouse Formation (Dickey and others, 1980; Turak, 2000). Trough and planar cross bedding as thick as 1 m and well sorted gravel and sand underlie and are interfingered with limestone at the base of the Bouse Formation. Limestone at a variety of elevations in Chemehuevi Valley contains mud cracks, mammal footprints, bird footprints (Reynolds, 2008) and gypsum suggesting a shallow, intermittently desiccated aquatic setting. Mud and sand overlie the limestone, except in some high-elevation areas where sand and angular gravel lenses overlie the limestone. This stratigraphic sequence is consistent with a model in which the arrival of clear Colorado River water was followed by slow filling of a large aquatic basin, first by limestone at the rising shoreline, then by mud and sand as the water body deepened. Local cross-bedded sand and mud balls at higher elevation in the formation may record wave action near the shoreline. Massive soft-sediment deformation structures in thick mud suggest syndepositional seismic activity. Bouse Formation limestone and sandstone locally are tilted in Chemehuevi Valley, along with overlying locally derived gravel beds. Elevations of highest preserved carbonate deposits in the formation decrease southward from Mohave Valley through Chemehuevi Valley to Parker Valley (cf. Spencer and others, 2008), in Chemehuevi Valley ranging from 305 to 335 m. Southward decrease may be the result of differential preservation, but also could be consistent with Lucchitta's (1979) inference of post-Miocene tilting of the Bouse Formation.

Kukla (1976) reported both normal and reversed magnetic polarity in the Bouse Formation. Near the Mesquite Mountains south of Parker, Arizona, our preliminary analysis of a 40-m-thick section of the formation reveals a paleomagnetic reversal at elevation ~500 m. Exploring for reversals at other sites could: (1) help constrain the possible chronology of the Bouse Formation and its rate of deposition, (2) potentially provide interbasin correlations of the formation, and (3) provide marker horizon(s) for measuring subsequent tectonic tilting.

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Quaternary Geology and Geomorphology of the Fremont River drainage Basin, South-central Utah

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The Fremont River drainage basin encompasses almost 2000 km^2 in south-central Utah and has a range of volcanic rocks and numerous glacial, colluvial, and alluvial deposits that provide excellent temporal markers for landscape evolution during the Cenozoic (figs. 1 and 2). The basin heads at the Basin and Range / Colorado Plateau transition zone in the High Plateaus of Utah. Late Oligocene to Pliocene aged volcanic rocks crop out in the western third of the drainage basin and provide information on the mid-Cenozoic history of the Colorado Plateau margin (Williams and Hackman, 1971). Multiple generations of normal faults have created a suite of grabens on the Fish Lake and Awapa Plateaus that provide information on the long range stress fields influencing the edge of the Colorado Plateau since late Miocene time (Bailey and others, 2007). The Fremont River flows across the last Basin and Range style normal fault, the Thousand Lakes Fault, at a place that G.K. Gilbert called Red Gate (Gilbert, 1877). To the north and south of Red Gate isolated patches of Oligocene aged volcanic rock (trachyandesite; ~25 Ma) cap Thousand Lakes and Boulder Mountains elevated as the footwall block of the Thousand Lakes fault. These mountains both sit above 3300 m with Boulder Mountain hosting an ice cap during the Pleistocene ice ages while Thousand Lakes Mountain was too small in area to support glacial ice. Both of the mountains are prone to mass movement and have generated rotational slumps, translational landslides, massive debris flows, and hyper-concentrated flow run outs (Billingsley and others, 1987). Mass movements from these two mountains have injected very coarse volcanic boulder debris into the main stem and tributary drainages of the Fremont River system.

The eastern 2/3rd of the Fremont River drainage basin is underlain by late Paleozoic to Mesozoic sedimentary rocks common to the central Colorado Plateau (Billingsley and others, 1987). Quaternary deposits in the eastern 2/3rd of the drainage basin almost always contain volcanic rocks, either from Boulder or Thousand Lakes Mountains (or smaller northern outliers of Hen Hole and Geyser Peaks) or from the Awapa or Fish Lake Plateaus carried by the main stem Fremont River. Although the volcanic rocks are generally intermediate in composition, they typically aquire very dark desert varnish and so are black in color, and contrast markedly with the red to white Mesozoic bedrock of the eastern part of the basin. Volcanic clasts in Quaternary (or perhaps older?) deposits are both physically and chemically more resistant to erosion that the local sedimentary bedrock. The strength of the volcanic clasts combined with the extremely large boulder sizes makes it difficult for local drainages to remove the coarse mass movement deposits. Over time, these deposits cause boulder armoring of pediments or straths and often lead to wholesale topographic inversion. There are likely more than a hundred of these topographically inverted, volcanic boulder-armored deposits around the Fremont River drainage basin and in the upper reaches of the Escalante River drainage basin along the southern slopes of Boulder Mountain (fig. 1).

The northern Henry Mountains make up a small portion of the far south-eastern part of the drainage basin. Porphyries in the Henry Mountains are close in age to the volcanic rocks capping Boulder and Thousand Lakes Mountains (Henry Mountains ~ 30–20 Ma, Boulder Mountain ~25 Ma: Mattox, 1991; Nelson and others, 1992) and are exposed at similar elevations. However, the porphyries in the Henry Mountains were likely emplaced at depth (perhaps 4 km deep, Nelson and Davidson, 1998), while the trachyandesites on Boulder and Thousand Lakes Mountains were deposited sub-aerially as massive ash flow sheets (Ball and others, 2009).

Quaternary deposits in the Fremont River drainage basin provide many promising targets for quantitative age dating attempts. Robert Anderson and students did some of the earliest work on cosmogenic ^{10}Be depth profile corrected exposure age dating of fluvial deposits in the lower Fremont River basin near Hanksville, UT in the early 1990's (Anderson and others, 1996; Repka and others, 1997) (fig. 1). They report inheritance corrected ^{10}Be ages of gravels capping three different straths of 60±9, 102±16, and 151±21 ka. From those data they estimate incision rates for the lower Fremont River of 0.30 to 0.85 m ka^{-1}.

The volcanic rocks that comprise most of the colluvial and alluvial deposits in the drainage basin have abundant pyroxene phenocrysts, which is an ideal mineral phase for cosmogenic ^{3}He exposure age dating. Our research group has taken advantage of these pyroxene-rich rocks for ^{3}He exposure age dating attempts throughout the basin (Marchetti and Cerling, 2005; Marchetti and others, 2005a,b; Marchetti and others, 2007). Published and unpublished ^{3}He exposure ages (likely minimum ages due to boulder erosion and deposit exhumation) of multiple boulders from many (~20 to date) of these mass movement and transitional flow run-out deposits range from 90 to 1200 ka, with individual deposit treads sitting 30 to 280 m above modern floodplains. These ages and incision depths (measured or estimated strath to strath incision depths) suggest maximum incision rates of 0.20 to 0.43 m ka^{-1} for the main stem Fremont River and its tributaries during the Quaternary.

The Quaternary deposits in the basin have well developed soils that frequently contain pedogenic carbonate and occasionally (generally in areas of the Moenkopi Formation), pedogenic gypsum. Pedogenic carbonate accumulations in some of the older soils are significant and many of the coarse gravelly deposits have pedogenic carbonate coatings (laminations, crusts) that are up to 8–10 cm thick. Although they vary in induration, stratigraphic continuity, and detrital content, many of these coatings provide excellent materials for U-series disequilibrium (^{230}Th/U) dating attempts (e.g. Marchetti and others, 2005b; see Sharp and others, 2003 for background).

The Fremont River drainage basin is an ideal natural laboratory to test many important hypotheses related to the overall incision of the Colorado Plateau and the geomorphic process of river incision in general.

- Quaternary deposits are abundant and at many heights around the basin. These deposits can be accurately dated using: ^{3}He on exposed boulder and desert pavements, ^{10}Be (or ^{21}Ne) on quartzite and chert clasts and possibly sandstones present in some of the deposits, U-series on pedogenic carbonates for deposits <500 ka, perhaps volcanic ash layers (e.g. older deposits may have thin layers of Lava Creek B or Bishop Tuff?), and some deposits may be amenable to cosmogenic burial or OSL dating.

- Since there are so many deposits, and because we can likely achieve well-constrained ages by utilizing a variety of dating techniques, this drainage basin could be used to directly test hypotheses related to changes in incision rate with time, and perhaps the passing of major knick-zones at times in the past.

- Most of the deposits in the basin are from mass movements or reworked mass movements. Therefore, the drainage basin should provide information on how mass movements affect incision rates and landscape evolution. This research could lead in interesting directions because mass movements may increase incision rates due to their strong tool effect and high shear stresses; however, after deposition they often cause significant bed armoring (cover effect) and may decrease fluvial incision and lead to drainage reorganization and topographic inversion (e.g. Sklar and Dietrich, 2001; Johnson and others, 2008; Cook and others, 2009).

- Finally, the glacial history of the drainage basin has been well studied and the extents and timing of middle and late Pleistocene glaciations on the Fish Lake Plateau and Boulder Mountain are known reasonably well (Hardy and Muessig, 1952; Flint and Denny, 1958; Marchetti and others, 2005a; Marchetti and others, *in press*).

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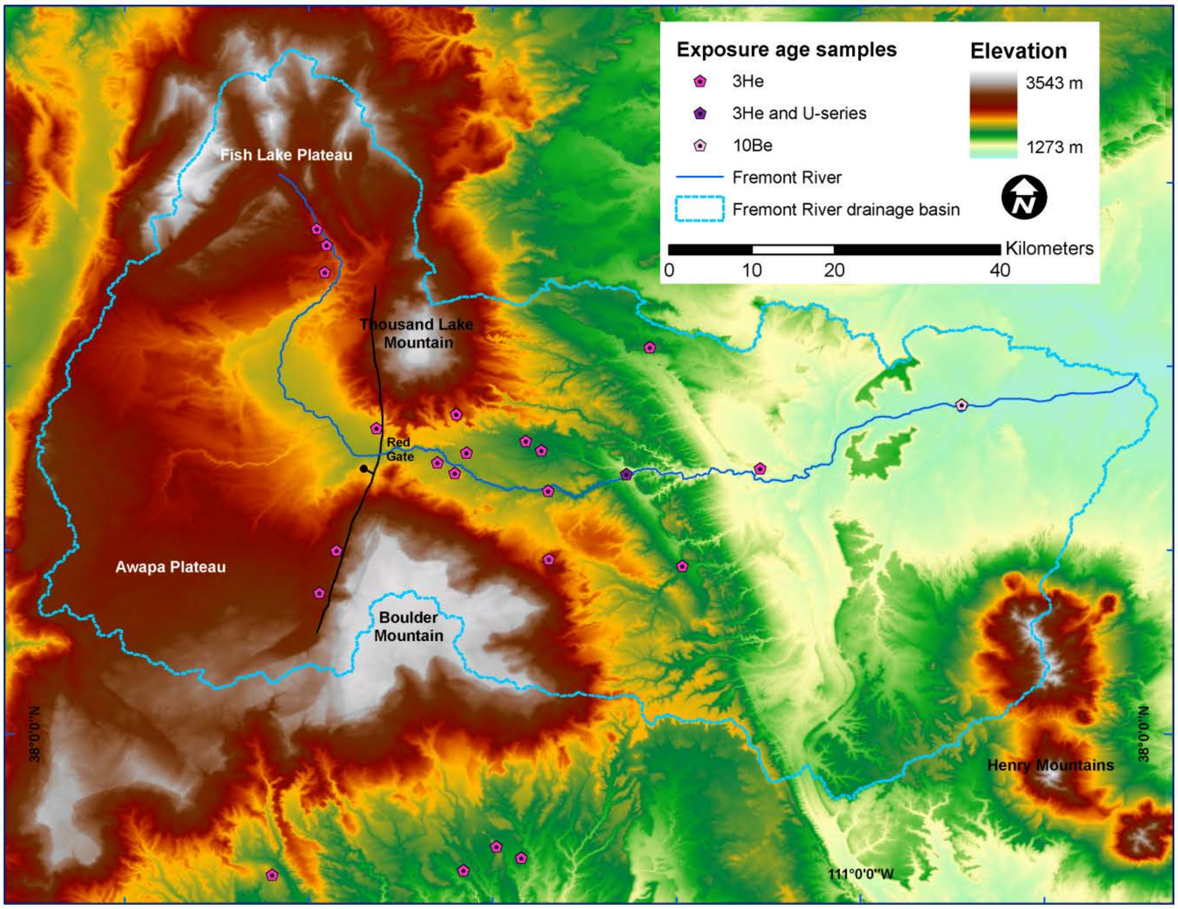

Figure 1. DEM of the Fremont River drainage basin. Locations with ^{3}He and ^{3}He and U-series ages are from Marchetti and Cerling, 2005; Marchetti and others, 2005b; and Marchetti and others, 2007. Several additional locations with unpublished ^{3}He age data are shown on the map. The location with ^{10}Be data is from Repka and others, 1997. Marchetti and others, 2005b includes some ^{3}He ages (un-corrected for exhumation or transport exposures) for the two oldest terraces in Repka and others, 1997.

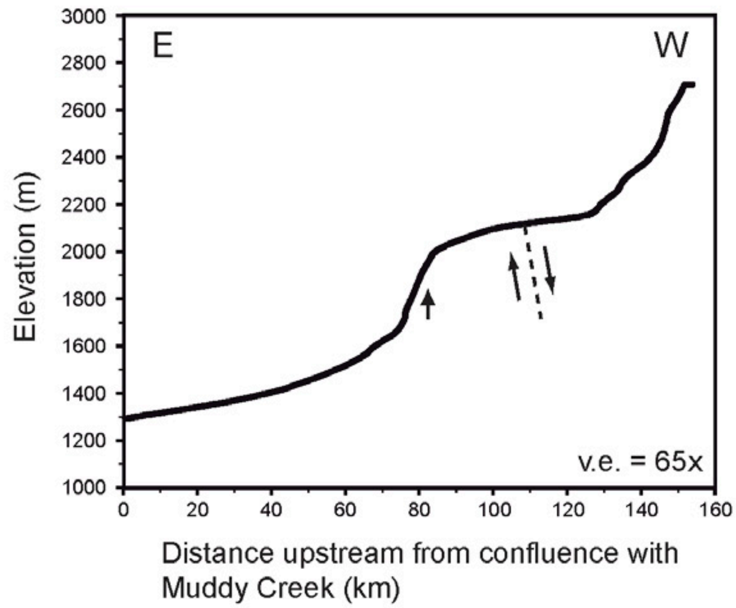

Figure 2. Longitudinal profile (valley profile) of the Fremont River. The large convexity in the profile is likely due to down-to-the-west slip on the Thousand Lakes fault (relative motion shown with arrows on dashed line representing fault plane) and resistant rock units exposed in the Fremont Gorge where the river crosses the axis of the Miners Mountain anticline (shown with single arrow).The profile starts at Johnson Valley Reservoir in the Fish Lake graben. Mill Meadow Reservoir was removed from the profile.

Geologic Evolution of the Mid-Tertiary Ash Creek Paleovalley, Black Hills, Central Arizona

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The Ash Creek Paleovalley of central Arizona is a northeast-southwest-trending exposure of Miocene and Oligocene(?) conglomerate that cuts across the top of the Black Hills, west of Camp Verde (Martin, 2003). In the paleovalley, the conglomerate depositionally overlies Proterozoic Grapevine Gulch Formation along an unconformity that records removal of most Paleozoic strata prior to the deposition of the conglomerate. The top of the Proterozoic rocks is lowest within the trough of the paleovalley and rises in elevation toward both the northwestern and southeastern flanks of the paleovalley. N ear the axis of the paleovalley, the basal contact of the conglomerate (the unconformity) slopes gently to the southwest, as does bedding in the conglomerate. The conglomerate and flanking Proterozoic rocks are overlain by basalt flows of the Miocene Hickey Formation. Near the axis of the paleovalley, the basal contact of the basalt flows dips gently to the southwest.

The conglomerate reaches a thickness of 70 m in the center of the paleovalley, near Ash Creek Well. It generally is poorly sorted and matrix-supported, but is locally clast rich (Sanders, 1989, Martin, 2003). The clasts range from pebbles to 2-m-diameter boulders and are subangular to subrounded. Clast types are diverse and include Grapevine Gulch Formation, gray feldspar porphyry, Mississippian Redwall Limestone, Devonian Martin Formation, chert, red granite, iron formation, Proterozoic Cherry Tonalite, gray fine-grained granite, Cambrian Tapeats Sandstone, green porphyry, metasedimentary rock, granodiorite with distinct biotite books, gray quartzofeldspathic metamorphosed rock, gabbro, Precambrian basalt, pink metarhyolite, milky vein quartz, calcite crystals, and basalt.

Clast imbrication in the conglomerate indicates paleoflow in the main northeast-trending part of the paleovalley was to the northeast, up the current dip of the beds. A small, southeast-trending, conglomerate-filled tributary to the paleovalley, exposed along present-day Tex Canyon, has a southeastern paleoflow. The main paleovalley probably joined other northeast-flowing drainages that transported Proterozoic clasts toward the Colorado Plateau. The modern drainage through Ash Creek flows to the southwest, requiring a drainage reversal since the deposition of the conglomerate. It is unknown if the drainage reversal recorded along Ash Creek occurred at the same time as the regional drainage reversal on the Colorado Plateau, but it shows the same pattern, switching from a northeastern to a southwestern flow direction. Hickey basalt flows dip toward the paleovalley from both flanks and indicate that the paleovalley was not completely filled, or had been partially exhumed, prior to eruption of the basalts. The overall southwestern dip of conglomerate beds, the basal unconformity, and basalt flows record post-conglomerate tectonic tilting, probably related to flank uplift associated with normal faulting that downdropped rocks to form the Verde Valley.

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Dating of Pliocene Colorado River Sediments—Implications for Cosmogenic Burial Dating and Evolution of the Colorado River

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We applied cosmogenic $^{26}Al/^{10}Be$ burial dating to sedimentary deposits deposited by the ancestral Colorado River downstream from Grand Canyon. All dated gravels yielded ages that suggest one or more episodes of sediment burial between ~5.3 and 3 Ma. Two minimum burial ages averaging roughly 4.6-Ma in the eastern Lake Mead area (Sandy Point, figs. 1, and 2) compare well to an independently dated overlying 4.4-Ma basalt (Faulds and others, 2001), and they suggest that under the most favorable conditions, cosmogenic burial dating is useful for deposits as old as 4–5 million years. Results highlight the complexities inherent in burial dating––complexities that arise from unknown and complicated burial histories, insufficient shielding, post-burial production of cosmogenic isotopes by muons, and unknown initial $^{26}Al/^{10}Be$ ratios. Nevertheless, and in spite of the large range of burial ages and large uncertainties, we can identify samples that provide reasonable burial-age constraints on the depositional history of sediment along the lower ancestral Colorado River. Our interpretations of sample data suggest deposition and burial 3.6±0.5 Ma for sediments perched high above Hoover Dam (Howard and others, 2008). A preliminary burial age ≥4.1±0.3 Ma at Topock in Mohave Valley for the alluvium of Bullhead City of House and others (2005) is consistent with 4.1±0.5 and >3.3 Ma tephra-correlation ages reported for that unit (House and others, 2008). A calculated burial age ~5.35+1.7/-1.0 Ma for the sediments of Hualapai Wash is consistent with their stratigraphic position as the earliest Colorado River sediment (Howard and others, 2008).

A basinwide erosion rate calculated from cosmogenic isotopes for modern sediment (collected at Needles, California) transported by the Colorado River (~187 mm ky^{-1}) is comparable to erosion rates inferred from the river's historical sediment load (~160 mm ky^{-1}), despite simplifying assumptions of source elevation and despite the enormous size of the drainage basin. In contrast, basinwide erosion rates calculated using the same assumptions from Pliocene river-laid sediments are all <50 mm ky^{-1} and offer a promising avenue for further research. Lower modeled rates for the Pliocene sediment samples are surprising given that the sampled time intervals include significant Pliocene aggradation and may include much incision of Grand Canyon and its tributaries. Possible reasons for lower Pliocene modeled rates may include extensive storage of sediment along the route of the Colorado River, slower paleobedrock erosion, or the inclusion of sediments that were derived preferentially from higher elevations in the watershed.

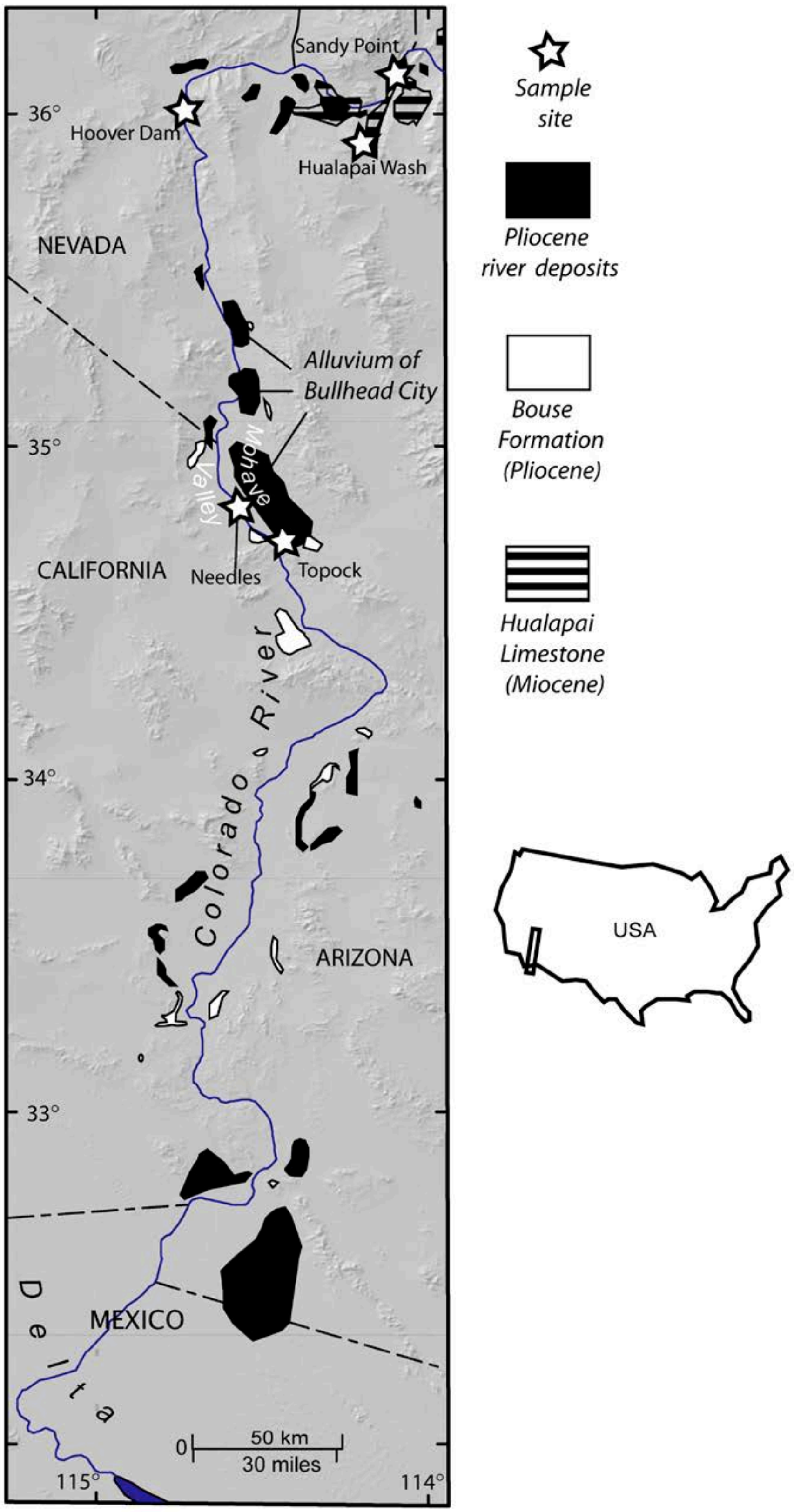

Figure 1. Sample localities along the valley of the lower Colorado River.

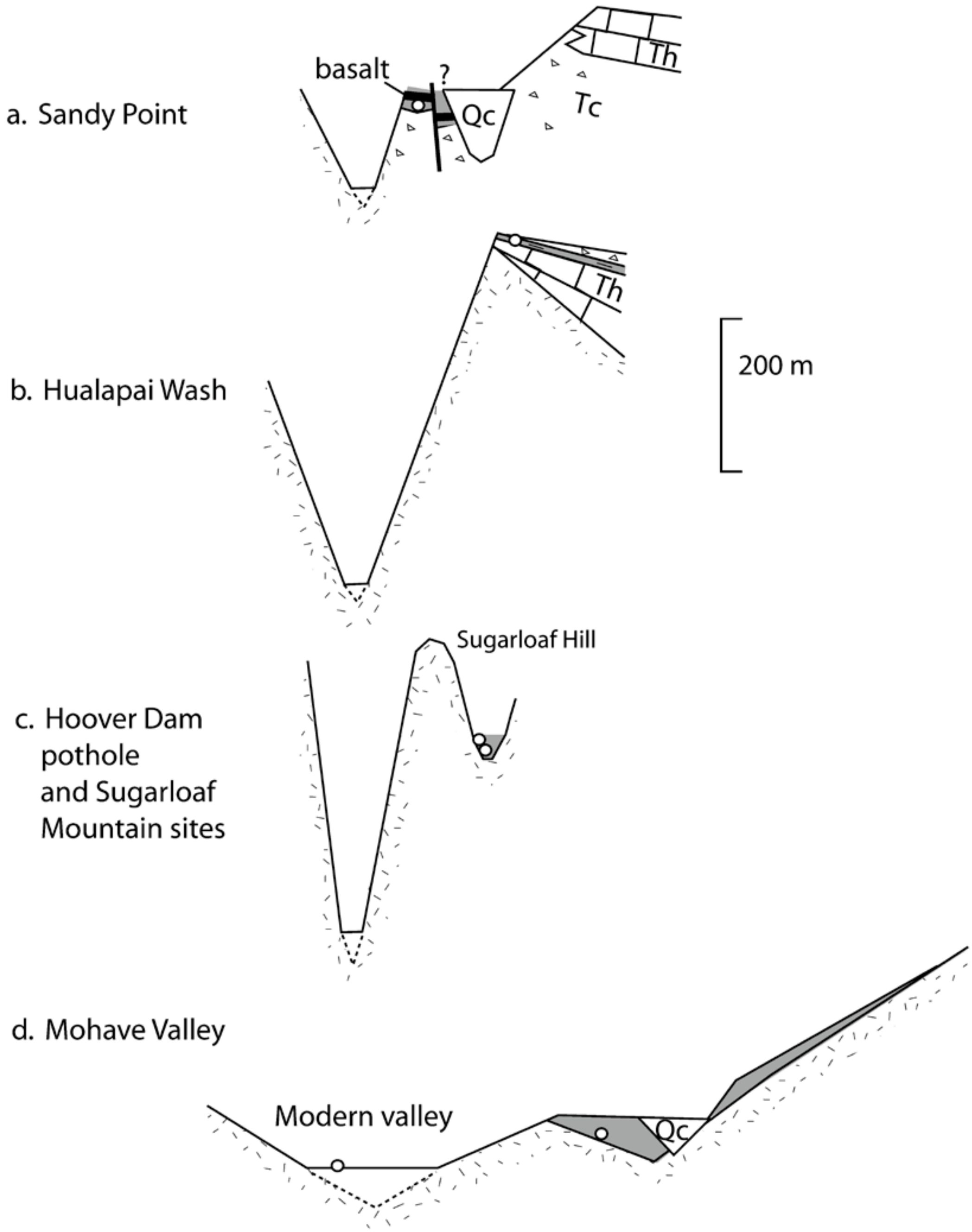

Figure 2. Diagrammatic sections of successively inset Colorado River deposits from upstream (A) to downstream (D), showing sample sites (blqck dots) in relation to Pliocene river deposits (shaded) and basalt (black), Miocene conglomerate (Tc), Miocene Hualapai Limestone (Th), upper Pleistocene Chemehuevi Formation (Qc; D. Malmon, written commun.), and alluvial fill (dashed) below the modern river valley. Pliocene and Miocene rocks are tilted at Sandy Point and Hualapai Wash. Deposits of Hualapai Wash predate Colorado River incision; other sampled units postdate some Colorado River incision.

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Update on Microfossil Studies in the Northern Gulf of California, Salton Trough, and Lower Colorado River

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My microfossil studies have focused on the opening of the Gulf of California, the age and extent of the subsequent marine incursions, and the arrival of the Colorado River (fig. 1). Marine sediments in the northern Gulf of California were deposited during marine incursions related to the opening of the Gulf through extension and strike-slip motion along major faults (Oskin and others, 2001; Oskin and Stock, 2003). Previous studies identified late Miocene sediments in the northern and central Salton Trough and along the lower Colorado River (McDougall and others, 1999; Dorsey and others, 2007; McDougall, 2008) and also recognized reworked middle Miocene microfossils in northern Salton Trough and northern Gulf sections, west of the San Andreas fault and East Pacific Ridge. These reworked middle Miocene microfossils occur in late Miocene sediments in the northern part of the Salton Trough, but are found in progressively younger sediments in the sections from the northern Gulf. Until recently, the existence of middle Miocene sediments has been questioned, but studies by Helenes and others (2009) of basins along the eastern side of the Gulf document middle Miocene marine sediments and faunas.

The distribution of middle Miocene sediments suggests that a proto-gulf lay along the eastern margin of the present Gulf. Marine waters extended as far north as the Altar Basin, and possibly to the Yuma area on the North American Plate. In place and reworked middle Miocene microfossils indicate that the proto-gulf was shallow but open-marine. The shallow-water interpretation is based on shelf benthic foraminifers, whereas the open-marine interpretation is based on planktic foraminifers and calcareous nannoplankton.

The distribution of late Miocene microfaunas indicate that tectonic activity changed from extension to strike-slip during late middle Miocene to early late Miocene time, based on the lateral and spatial offset of reworked middle Miocene faunas from their source areas. Marine waters extended as far north as the northern Salton Trough. Interpretation of late Miocene marine microfossils indicates tropical conditions and increasing water depths. The Miocene-Pliocene boundary is marked by an unconformity and a rapid increase in water depth (Dorsey and others, 2007), which allowed marine waters to inundate the Blythe Basin along the lower Colorado River.

The marine or nonmarine origin of the Bouse Formation is frequently debated, with some suggesting marine fossils were transported by birds (Spencer and Patchett, 1997). However, in the Blythe Basin, which stretches along the Colorado River from the Chocolate Mountains to Parker, Arizona, the marine and nonmarine parts of the Bouse Formation interfinger. Benthic foraminiferal species in the Blythe Basin contain a mixture of inner neritic (<50 m) and middle neritic (50–100 m) species. Not all of these species could have been transported by birds, but once established these assemblages could survive in a saline lake until salinity exceeded species tolerances. Planktic foraminifers also occur in the Bouse Formation of the Blythe Basin. Planktic foraminifers need open marine conditions to survive and tolerate only a very narrow salinity range. Avian transport of this group would be difficult. Planktic foraminifers are restricted to two intervals in the lower part of the section in outcrop and core samples, which are earliest Pliocene in age. Both occurrences of planktic foraminifers are at elevations below the estimated late Miocene and early Pliocene sea levels, assuming no change in elevation (fig. 2). The rapid deepening of marine-water depths (to approximately 300 m) near the Miocene-Pliocene boundary observed in the Salton Trough may coincide with at least one of these marine pulses.

Pliocene and Pleistocene microfossil assemblages in the Salton Trough are less diverse and are increasingly diagnostic of marginal marine conditions and shallow water. Influx of Colorado River water and sediments is suggested by the disappearance of tropical species and species adapted to clear water in the Pliocene. Reworked Cretaceous foraminifers from the Colorado Plateau also are present in some assemblages.

Although understanding of the Miocene and Pliocene microfossils in sediments from the Salton Trough and along the lower Colorado River has improved, more work is needed to document the connection between the proto-gulf, Salton Trough, and lower Colorado River areas and to refine the timing of the events.

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Figure 1. Index map of study area showing geographic and geologic features and sedimentary basins discussed in the text. The base map is modified from the U.S. National Park Service (NPS) Natural Earth physical map (*http://goto.arcgisonline.com/maps/World_Physical_Map*).

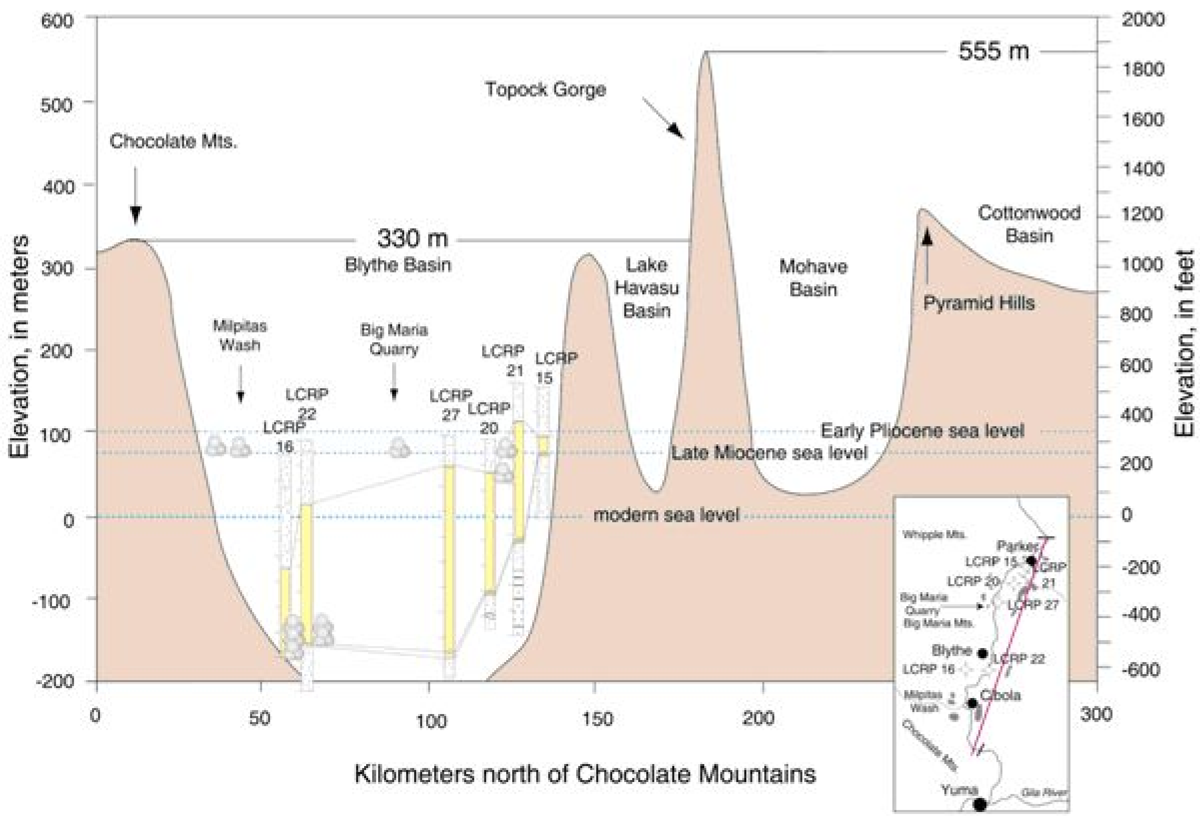

Figure 2. Cross-section of the Blythe Basin showing location of wells, outcrop sections, planktic foraminifers, and sea-level elevations. Data from Smith (1970) and McDougall, unpublished. Cross-section adapted from Spencer and others (2008). Red line on insert map shows extent of Blythe Basin.

Ancestral Colorado River Exit from the Plateau Province—Salt River Hypothesis

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Earlier workers proposed a westward-flowing ancestral Colorado River that may have originated as early as Oligocene time, yet little is known of its exit from the Colorado Plateau prior to Pliocene time. Aslan and others (2011) confirm the existence of westward flowing streams from the Rocky Mountains prior to 11 Ma. At the mouth of Grand Canyon, the Muddy Creek Formation composed of locally derived gravels precludes a region-wide southwest flowing river system through Grand Canyon until after ca. 5-6 Ma (House and others, 2005). Alternative river exit points from the plateau province have been proposed but are shown to be unlikely (Pederson, 2001). In this extended abstract several lines of evidence support an exit point for the ancestral Colorado River at the Mogollon Rim in eastern Arizona, which would then have followed the upper Salt River (White River) into the Gila River system of southern Arizona.

Bidahochi Formation stratigraphy and sedimentology

The Bidahochi Formation provides a sedimentary record of a Miocene lacustrine- volcanic-fluvial system on the southern Colorado Plateau (fig. 1). The lower lacustrine member was deposited in the valley of the Little Colorado River between ca. 15.8 and 13.7 Ma followed by volcanism of the middle member between ca. 8.5 and 6.6 Ma (Dallege and others, 2001). Late Miocene/Pliocene fluvial deposits of the upper member rise in elevation and overstep the lower members toward the southeast (Love, 1989), suggesting progradation and shift of the depocenter toward the eastern Mogollon Rim (fig. 1). Given the paucity of evaporites in the Bidahochi Formation, this depositional basin required an outlet to prevent high salinity for about nine million years. As such, the Bidahochi basin is here interpreted as a broad shallow pass-through basin that ponded sediment and surface water en route to a downstream destination.

Proximity of the Bidahochi Formation to the eastern Mogollon Rim

Proximity to the Mogollon Rim and present altitudes of the Bidahochi Formation allow for possible overtopping of the Mogollon Rim in eastern Arizona (fig. 2). Today, the base altitude of the upper fluvial member nearest to the Mogollon Rim is 2200 m near the town of Springerville (Love, 1989) 60 km northeast of the postulated exit point along the Mogollon Rim (fig. 3). The top altitude would be about 2250 m, assuming a 50 m thickness of the upper member (Dallege and others, 2001). At Amos Mountain on the Mogollon Rim, two distal Mount Baldy volcanic flows overlie deeply eroded Mogollon Rim Formation at elevation 2200 m (Condit, 1991) indicating pre-Baldy erosion of 300+ m of volcanic and sedimentary rocks by an unknown stream system following emplacement of the Lower Miocene White Mountain volcanic field (fig. 3). This predecessor stream (ancestral Colorado River?) was active during a period bracketed between the youngest White Mountain volcanic rocks (ca. 21 Ma) and the basal Mount Baldy flow on Amos Mountain (8.97 Ma). This time frame is consistent with rock ages obtained for the lower member of the Bidahochi Formation.

Northward flexure of the southern Colorado Plateau in the Neogene

Rift shoulder uplift of the Colorado Plateau's southeastern margin during Basin and Range/Rio Grande rifting might be expected as is observed today on the flanks of the East African Rift. The eastern Mogollon Rim is the southerly crest of a broad northeast-dipping homocline of Paleozoic/Mesozoic strata called the Mogollon Plateau (Richard and others, 2002). Northeastward flexure of the 18.6 Ma Apache Leap tuff mapped on the southwest flank of the Salt River paleocanyon 50 km south of the Mogollon Rim (fig. 5) demonstrates flexure during the Neogene. North of the Mogollon Rim, elevation gradients of the basal Bidahochi Formation indicate post-Miocene northward and westward flexure of the southeastern Colorado Plateau and Bidahochi basin. The base of the Bidahochi Formation gains 450 m of elevation from depocenter to the east and south (Love, 1989, fig. 1) suggesting flexure of the basal Bidahochi Formation following deposition. Assuming no Neogene flexure of the southern Colorado Plateau margin, the Bidahochi Formation would be required to onlap a 450 m highland to the east and south, improbable given its average total thickness of 190-200 m (Dallege and others, 2001).

Progressive incision history of the upper Salt River during Neogene time

Southwest of the Mogollon Rim and east of Canyon Creek Fault volcanic flow remnants along the flanks of the White River valley (upper Salt River) delineate a southwest-directed incision history that spans much of Neogene and Quaternary time. Fluvial incision began sometime after emplacement of the extensive White Mountain volcanic field ca. 27-21 Ma (Berry, 1976, Potochnik, 1989). The Mount Baldy volcano complex, which straddles the Mogollon Rim, was emplaced primarily on a southerly paleoslope (Merrill and Pewe, 1976). Remnants of Mount Baldy volcanic flows are inset beneath and against the incised Mogollon Rim Formation/White Mountain Volcanic Field (fig. 4) at progressively lower elevations southward along the flanks of the White River valley indicating that southwest stream flow began prior to ca. 9 Ma (figs. 3, 4). At Nan Dahs Taan mesa adjacent to the town of Whiteriver, a yet lower elevation basalt flow overlies a thin remnant of the arkosic Mogollon Rim Formation, indicating erosion of 375 m of late Laramide fluvial clastics by a southwest-flowing stream prior to 3.5 Ma (fig. 4). Lastly, Pleistocene basalt flows from the Springerville Volcanic Field line the walls of the modern White River and its tributaries, demonstrating that the White River had nearly attained its present grade prior to ca. 1.87 Ma (Condit, 1991) (figs. 3 and 4). These rock relationships delineate a progressively deeper incision of the upper Salt River valley by a southwest flowing stream that originated on the Colorado Plateau between 21 Ma and 9 Ma, coeval with deposition of the Bidahochi Formation.

Stratigraphy and geomorphic history of the Salt River paleocanyon

West of Canyon Creek Fault (fig. 2), a paleocanyon more than 1000 m deep was incised into Proterozoic and Paleozoic bedrock by a northeast-flowing Laramide stream system (fig. 5). Oligocene and Miocene deposition of more than 300 m of northeast-transported Whitetail Conglomerate stream gravels within this paleocanyon concluded with the emplacement of 14.84 Ma Black Mesa basalt flows (Potochnik and Faulds, 1998). Following drainage reversal, a subsequent paleocanyon was then incised at least 300 m into the Whitetail Conglomerate by a southwest flowing ancestral river after 14.84 Ma. This was followed by deposition in this Miocene inset paleocanyon of at least 313 m of Dagger Canyon conglomerate by the same river. The modern Salt River has, in turn, incised this entire sequence of inset stratigraphic units (Potochnik, 2001a, b). Error bars on the earliest age date of the Bidahochi Formation (15.46+/-0.58 Ma) and on the Black Mesa basalt (14.84+/-0.43 Ma) allow for the possibility that Lake

Hopi established an outlet through this ancestral Miocene canyon, providing a conduit for ancestral Bidahochi Formation water and sediment southwestward toward the nascent Gila River trough during onset of Basin and Range extension.

Basin and Range repositories of Neogene sediment and water

Toward the close of northeastward stream flow in the Early-Mid Miocene the topographic difference between Colorado Plateau and Basin and Range would likely have been relatively small. Widespread deposition of submature Whitetail Conglomerate in the Transition Zone during the Oligocene and Miocene speaks to a long period of increasingly incompetent northeastward stream flow in central Arizona, ending with local infilling and lacustrine deposition as seen in the Great Basin of today (Potochnik and Faulds, 1998). The gentle southward sloping land surface beneath the Mount Baldy volcanic complex indicates that the exit point for pre-Baldy drainage along the Mogollon Rim was topographically subtle during the Middle Miocene phase of Basin and Range subsidence (Merrill and Pewe, 1977). Late Miocene to Pliocene Basin and Range crustal thinning and subsidence south of the Colorado Plateau progressively increased the topographic relief and base level fall for southwest flowing streams. The Tonto, Luke, and Higley basins would have provided large repositories for sediment as Miocene streams flowed southwestward toward the Gila trough. In the Luke basin alone, at least 1300 m of evaporites were precipitated during the Basin and Range event by an unknown river (Spencer and Rauzi, 2005) mostly prior to emplacement of basalt flows ca. 10.5 Ma (Shafiqullah and others, 1980). The Luke and Higley basins are among the deepest in the southern Basin and Range, shown by gravity surveys to reach bedrock depths exceeding 3600 m (Oppenheimer and Sumner, 1980). If these basins overflowed southwestward down the emerging Gila trough, this could explain the distinctive detrital zircon signature of the lower Gila River (Kimbrough and others, 2011). This would reflect extensive exhumation of the 1.4 Ga Ruin Granite in the Salt River paleocanyon and subsequent recycling of resultant arkosic sand from the late Eocene Mogollon Rim Formation.

Salt River outlet from the Colorado Plateau is closed, which forces surface runoff toward the Grand Canyon outlet

This postulated exit for the ancestral Colorado River was initially obstructed by emplacement of the late Miocene Mount Baldy latite flows ca. 9 Ma, which although restricted in lateral extent, reduced stream flow from the Bidahochi basin and may have induced aggradation of the Dagger Canyon conglomerate in the ancestral Salt River Canyon (Potochnik 2001a). Additionally, northward flexure of the Mogollon Plateau may have caused final obstruction of ancestral Colorado River streams. Some combination of volcanic blockage and flexure would have forced ancestral Colorado River runoff to find an alternative exit from the Bidahochi basin and the Colorado Plateau.

The topographically lowest alternative exit point may have been Cape Solitude (1873 m) overlooking the confluence of the Little Colorado and Colorado River. Here, the Kaibab Formation is overlain by a limey marl of unknown age on the south rim of Grand Canyon (Scarborough, 2001). Given the profound bedrock canyon cutting history of the Salt River during the Laramide Orogeny, it is conceivable that an analogous Laramide canyon may have been incised by a northeastward-flowing stream to the level of Cape Solitude through the Kaibab Plateau (Flowers and others, 2008). If so, then a convenient outlet would be provided for the

Pliocene Colorado River to exit through a pre-existing ancestral Grand Canyon (Potochnik, 2001b) while the earlier Salt River outlet was further blocked by Plio-Pleistocene flood basalts of the Springerville Volcanic Field.

Summary of hypothesis

In this paleogeographic model, the nascent ancestral Colorado River was ponded in the Bidahochi Basin then, overflowed the southern edge of the Colorado Plateau in eastern Arizona during incipient Basin and Range faulting about 15 Ma. This early southwest- flowing stream utilized pre-existing Laramide topography as it followed paleotopographic lows across the Transition Zone province, ponding in newly developing structural troughs of the southern Basin and Range Province. Volcanic and/or flexural blockage of the ancestral Colorado River on the Mogollon Rim then forced the ancestral Colorado River westward to overtop the Kaibab Formation at the confluence of the Little Colorado and Colorado Rivers into a pre-existing Laramide paleocanyon. This diverted the principle drainage from the Salt River outlet toward the Grand Canyon outlet, incising the Muddy Creek Formation and delivering sediment/water load of the modern Colorado River to the Pliocene Salton trough.

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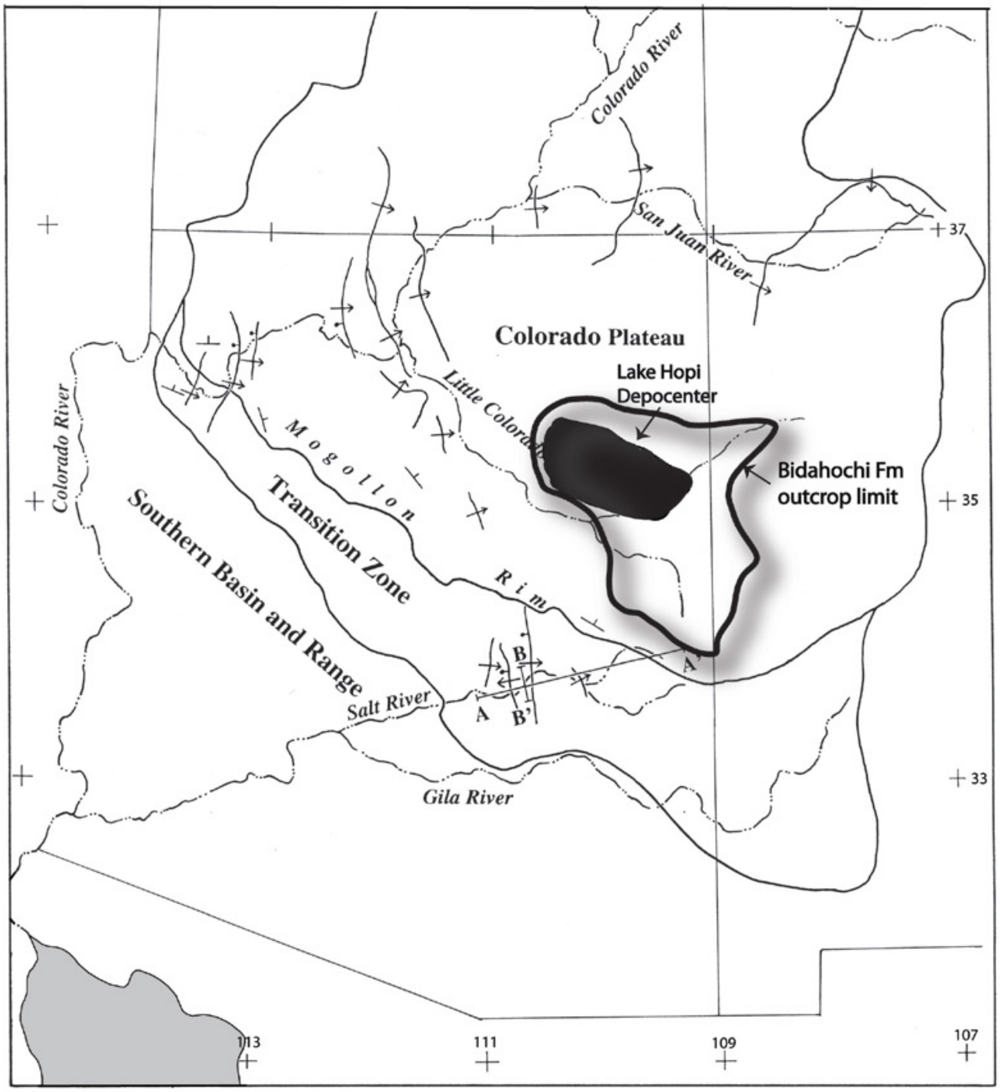

Figure 1. Note proximity of Lake Hopi depocenter and progradation of Bidahochi Formation toward headwaters of the Salt River near the Mogollon Rim of eastern Arizona (after Potochnik, 2001b).

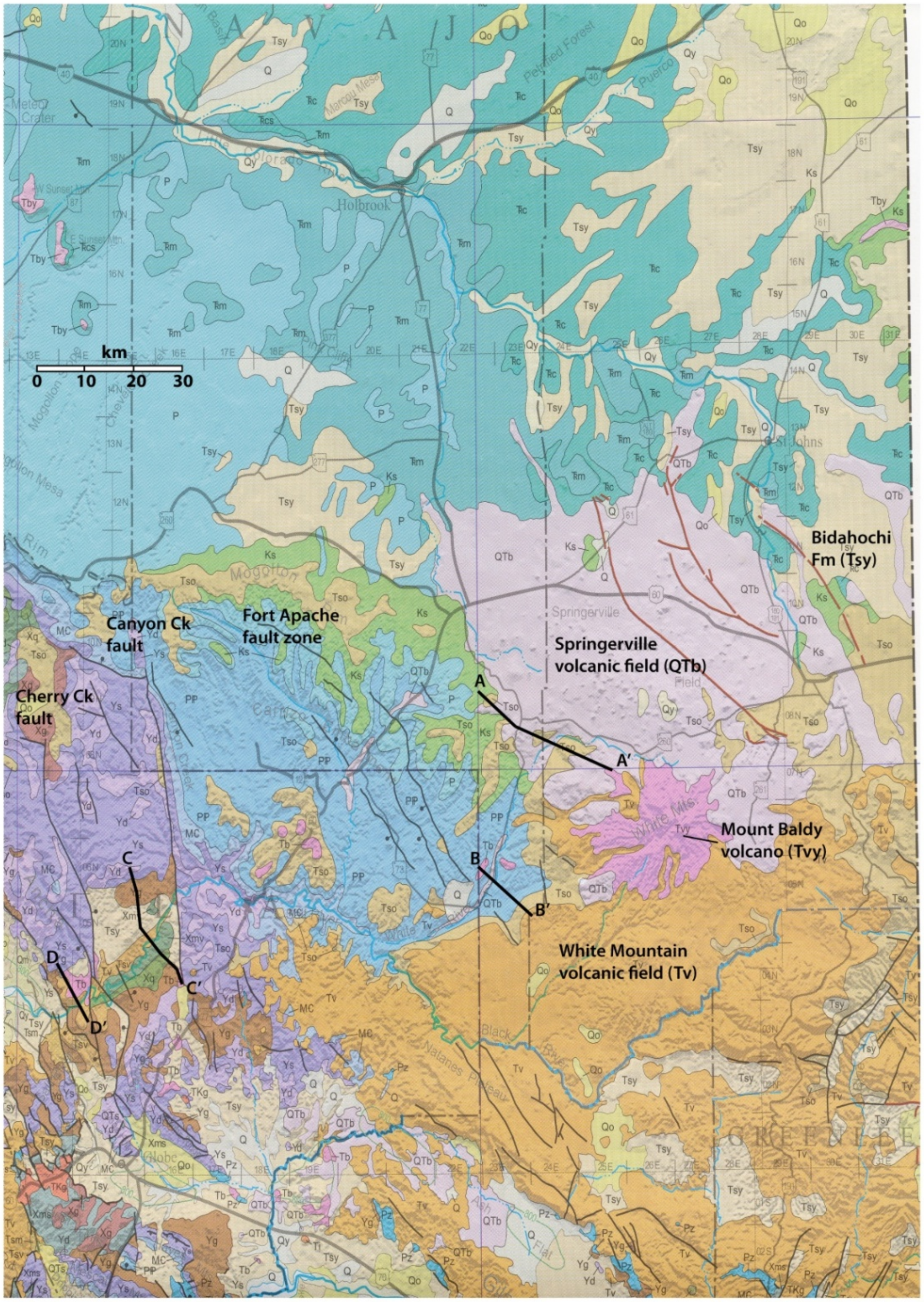

Figure 2. Geologic map of eastern Mogollon Rim and the "Carrizo embayment", an erosional escarpment of the upper Salt River defined by upper Paleozoic rocks to the north and lower Miocene volcanic flows of the White Mountain Volcanic Field to the east (after Richard and others, 2002). Locations are shown of cross-sections in figures 3-5. Lithic designators on this Arizona State map are not used in the other figures in this abstract.

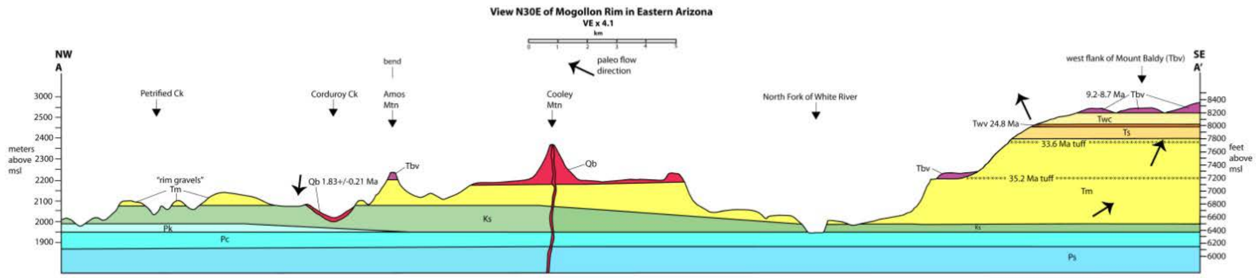

Figure 3. Cross section A-A'. Parallels the Mogollon Rim where Paleozoic sedimentary rocks transition eastward to edge of White Mountain Volcanic Field and Mount Baldy volcanic complex. Mount Baldy distal flow remnants that cap Amos Mountain are the western-most exposures of these ca.8-9 Ma volcanic rocks. These volcanics constrain at least 300 m of erosion of the earlier deposited Mogollon Rim Formation and overlying volcanics and volcaniclastics of the White Mountain volcanic province by an earlier stream. Early Pleistocene Springerville volcanics flowed southward down modern drainages of Corduroy Creek and North Fork of White River.

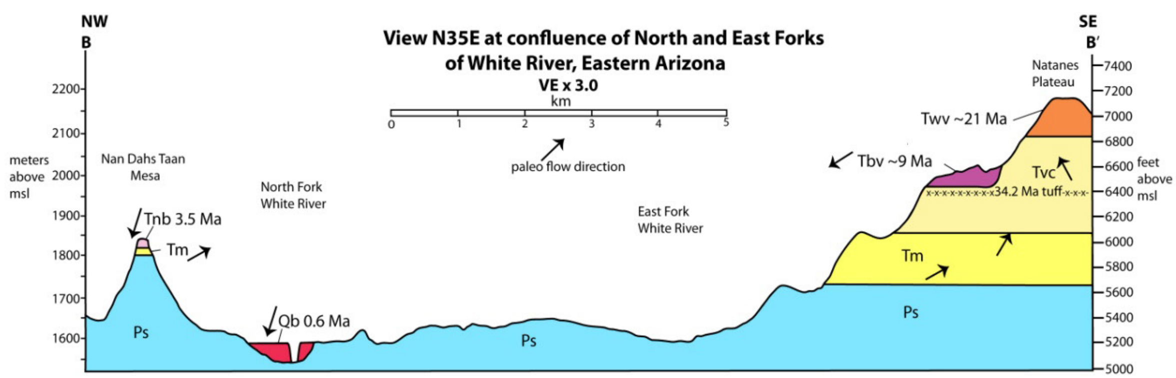

Figure 4. Cross section B-B'. At confluence of North Fork and East Fork of White River, progressive incision history is preserved beneath volcanic flows of Oligocene to Pleistocene age. Most of the earlier deposited arkosic sands of the Mogollon Rim Formation were reworked southwestward during excavation of this valley following drainage reversal. Note progressive erosion loss of Mogollon Rim Formation beneath younger volcanic flows.

Figure 5a.

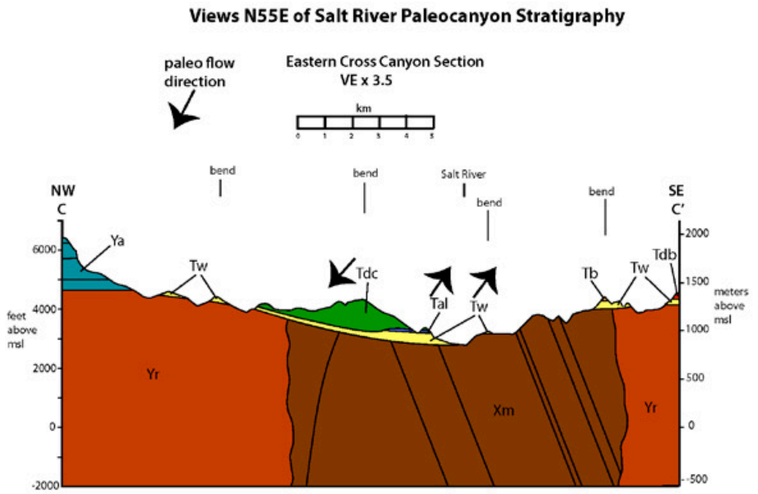

Figure 5b.

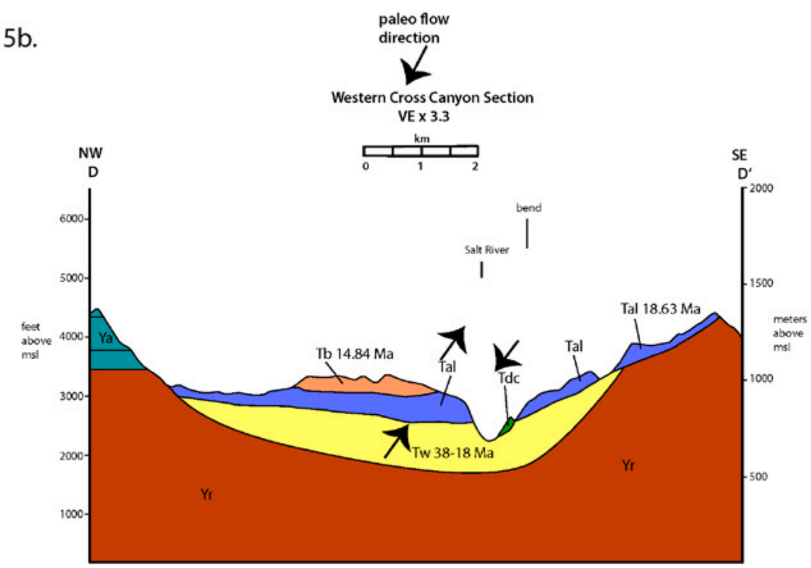

Figure 5. Cross sections C-C', D-D'. Two cross sections show different aspects of the Salt River paleocanyon of Laramide age, later occupied by the postulated ancestral Colorado and modern Salt Rivers (after Potochnik, 2001a, b). Bedrock Laramide paleocanyon was carved by a northeast flowing stream more than a thousand meters into Proterozoic bedrock during Mogollon Highland uplift. Laramide topography governed the course of later ancestral Colorado and modern Salt River following drainage reversal.

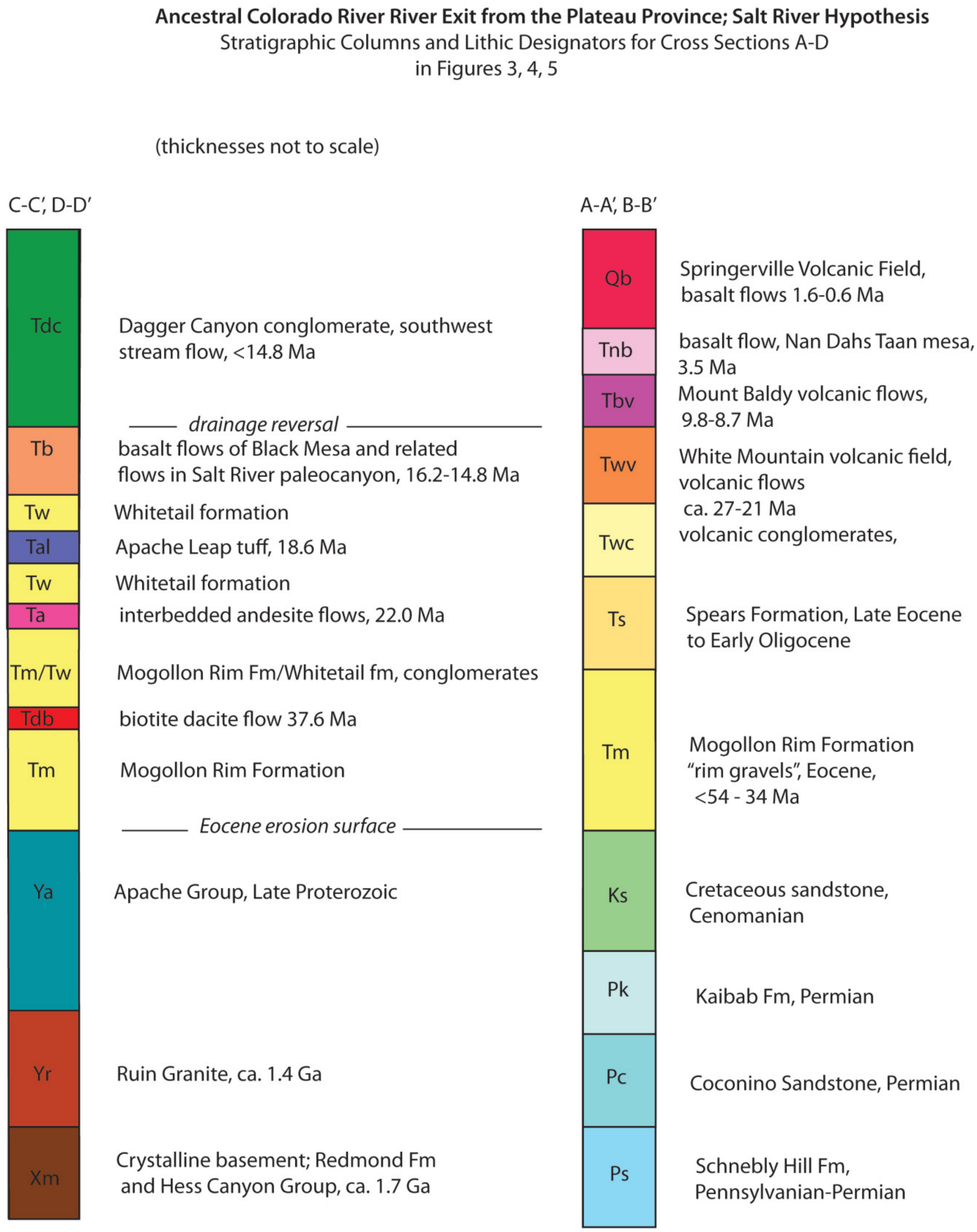

Figure 6. Stratigraphic columns east and west of Canyon Creek Fault show map units and stratigraphic order for rock formations as used on cross sections in figures 3-5. Thickness of rock formations is not to scale.

A Tale of Two Monoclines

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Monoclines of the western Grand Canyon

The transition between the Colorado Plateau and Basin and Range tectonic provinces (fig. 1A), exposed in the western Grand Canyon and Lake Mead regions, provides an excellent natural laboratory for studying processes of crustal extension. One of the striking features of the fault systems within this transition zone is the occurrence of relatively narrow (2-10 km) hanging wall monoclines along many fault segments. These folds might be more properly termed half-monoclines, with bed dips increasing toward the associated normal faults. These (half) monoclines of the western Grand Canyon inspired the most widely cited paper on the development of these "reverse drag" folds, wherein Hamblin (1965) proposed that the folds formed in response to a listric fault geometry. Many mysteries, however, remain regarding the processes that create these folds. Even their direct association with normal faulting has been called into question (Huntoon, 1990, 2003). In this abstract we describe two of these monoclines in detail: The Hualapai Cove monocline (new name) associated with the Wheeler and Lost Basin Range faults in the Lake Mead region and the Lone Mountain monocline associated with the Frog Fault in the western Grand Canyon. Although these faults and folds cut and deform different bedrock types at the surface, the shape and width of the folds is similar to a first order, an observation that leads us to look for a common physical process leading to the formation of narrow reverse drag folds.

Wheeler/Lost Basin Range Faults and folding of the Hualapai Limestone

The 12-6 Ma Hualapai Limestone (Faulds and others, 2001; Spencer and others, 2001; Lopez Pearce and others, 2010 and this volume) is down-dropped more than 275 m across the Wheeler and Lost Basin Range faults (Wallace and others, 2005). This offset records low-magnitude post-Middle Miocene extension across the western margin of the Colorado Plateau and eastern Basin and Range that appears to continue to this day (Kreemer and others, 2010). A newly discovered exposure of the Lost Basin Range Fault surface strikes 202°, dips 55° west, and has slickenlines with a rake of 76° SW. The dip of this fault surface is consistent with previous estimates of fault dip from 3-point solutions (Wallace and others, 2005). Along much of this fault system a portion of the displacement is accommodated by down-warping of the Hualapai Limestone into a hanging wall monocline (fig. 1B). This monocline is best developed in the vicinity of Hualapai Cove and we therefore refer to the structure as the Hualapai Cove monocline. The monocline warps the Hualapai Limestone from an elevation of >700 m east of Little Burro Bay (~ 5 km from the fault) to an elevation of ~400 m in the trough of the Gregg Basin syncline. Beds dip 10-15° east across much of the exposed fold with maximum eastward dips reaching 22° east of Hualapai Wash. Fold geometry varies along strike with down-warping (reverse drag) dominating along much of the Lost Basin Range fault, giving way to a broad hanging-wall (Gregg Basin) syncline at the fault's northern end, a structure we interpret to have formed in response to propagation of the fault upward through previously deposited Hualapai strata. The Wheeler fault exhibits minor reverse drag (east dips to 6°) as well as short wavelength “normal drag” folds (west dips to 30°). In the relay zone between the Lost Basin and Wheeler faults beds are tilted northward in a ramp geometry. Near-fault footwall exposure is limited to the northern end of the study area and is complicated by a series of small-offset faults.

Frog Fault and Lone Mountain monocline

The Frog fault system and Lone Mountain monocline were described in detail by Resor (2008). We briefly review the main features of the system here. The Frog fault offsets Paleozoic strata of the Colorado Plateau up to ~225 meters. The Frog fault dips ~70° to the west and slip and is nearly pure dip-slip, as evidenced by slickenline orientations. The fault system is comprised of the Frog fault and a series of smaller-offset (secondary) synthetic and antithetic normal faults. A system of folds, including an upper half monocline in the hanging wall (Lone Mountain monocline of Huntoon and Billingsley (1981)) and a lower half-monocline in the footwall, parallels the Frog fault system (fig. 1C). The dip of hanging wall beds increases systematically toward the fault over ~1.5 km generating ~200 m of structural relief at the southeast end of the map area. This folding is largely absent at the northwest end of the map area where the majority of the faults within the system terminate. The dip of footwall beds decreases away from the fault over a distance of ~0.5 km with structural relief typically less than 25 meters. The net structural relief of the upper Esplanade across the Frog fault and associated monoclines is thus less than 25 m. Fold geometry changes along strike and appears to have a strong association with the development of secondary faults. The southeastern portion of the study area has the highest structural relief and greatest bed dips. In this area the fault system is composed of the main Frog fault with a ~3-km long synthetic fault in the hanging wall. The structural pattern changes dramatically to the northwest of Parashant Canyon. This portion of the Frog fault system is composed of two asymmetric grabens with the larger faults dipping toward the west. With the exception of two well exposed relay ramps that have steeper northwesterly dips, bedding dips are consistently shallower in this part of the fault system with maximum eastward dips of ~ 6°.

Modeling reverse drag

Since Laubscher (1956) and Hamblin (1965) first proposed a direct relationship between folding of hanging wall strata (reverse drag) and the geometry of underlying normal faults, many geologists have developed and employed geometric methods to relate fold and fault geometry. These “kinematic” methods require a listric (concave-upward) fault shape in order to generate reverse-drag folds. Others have noted, however, that similar hanging wall folds are predicted by mechanical models of planar faults that extend to a finite depth in the crust. These end-member models lead to significantly different interpretations of not only the subsurface fault geometry, but also the resulting tectonic strain. We employ a boundary element method (BEM) based on the solution of a two-dimensional dislocation discontinuity within an elastic half space (TWODD, Crouch and Starfield, 1983) to directly compare patterns of displacement around planar and listric faults and evaluate criteria that may be used to determine subsurface fault geometry from observations of near-surface deformation.

Models that incorporate a) a planar fault dipping 60 degrees with constant slip, b) a planar fault dipping 60 degrees with constant slip ending at a free-slipping horizontal detachment, and c) a listric fault dipping 60 degrees at the surface with constant slip ending at a free-slipping detachment all develop hanging wall reverse-drag folds whose width increases only slightly with introduction of the detachment and listric fault shape (fig. 2A). The most notable difference between listric and planar models may be the magnitude of footwall uplift (fig. 2B). Footwall uplift decreases slightly with introduction of the detachment and more significantly with the addition of a listric fault shape. A parametric investigation of faults with constant slip ranging from nearly planar to strongly listric over depths from 1 to 15 km (fig. 2C) reveals that hanging wall fold width is very sensitive to fault depth with little sensitivity to fault geometry. Footwall fold width and the ratio of footwall uplift to hanging wall subsidence, however, are sensitive to fault geometry. A combination of these observable features may therefore provide a tool to estimate fault geometry rather than simply inferring listric geometry based on the presence of reverse-drag folds.

Speculation on the development of reverse drag

Mechanical models of narrow reverse drag folds, such as the Hualapai Cove and Lone Mountain monoclines, are consistent with a relatively shallow source (1-5 km to the lower fault tip or detachment). In the case of the Frog fault the Bright Angel Shale at the basement-cover contact would seem a likely detachment or fault termination zone. In the case of the Wheeler fault, however, the 170-300 meter thick Hualapai Limestone overlies an irregular substrate of Miocene conglomerates, tilted Paleozoic strata, and Proterozoic crystalline basement (Wallace and others, 2005). The development of a shallow through-going detachment in this situation appears less likely. Furthermore, the significant variation in along-strike fold development highlights the importance of fault interactions and fault system geometry in the development of significant reverse drag. In the case of the Lone Mountain monocline there is a clear association between the magnitude of folding and the occurrence of synthetic faulting. In the Wheeler fault system overlapping faults in the relay zone may play a similar role in promoting growth of the Hualapai Cove monocline. A third consideration in the development of narrow reverse drag folds is the process by which elastic stresses are made permanent in the upper crust. Small scale fracturing, a likely mechanism for inelastic strain accommodation, may be more likely to occur in proximity to active fault zones. These areas may therefore accommodate greater long-term folding strains, localizing folding in a relatively narrow zone close to the fault. Each of these hypotheses warrants further examination and the Grand Canyon may be the best natural laboratory we have to study these processes.

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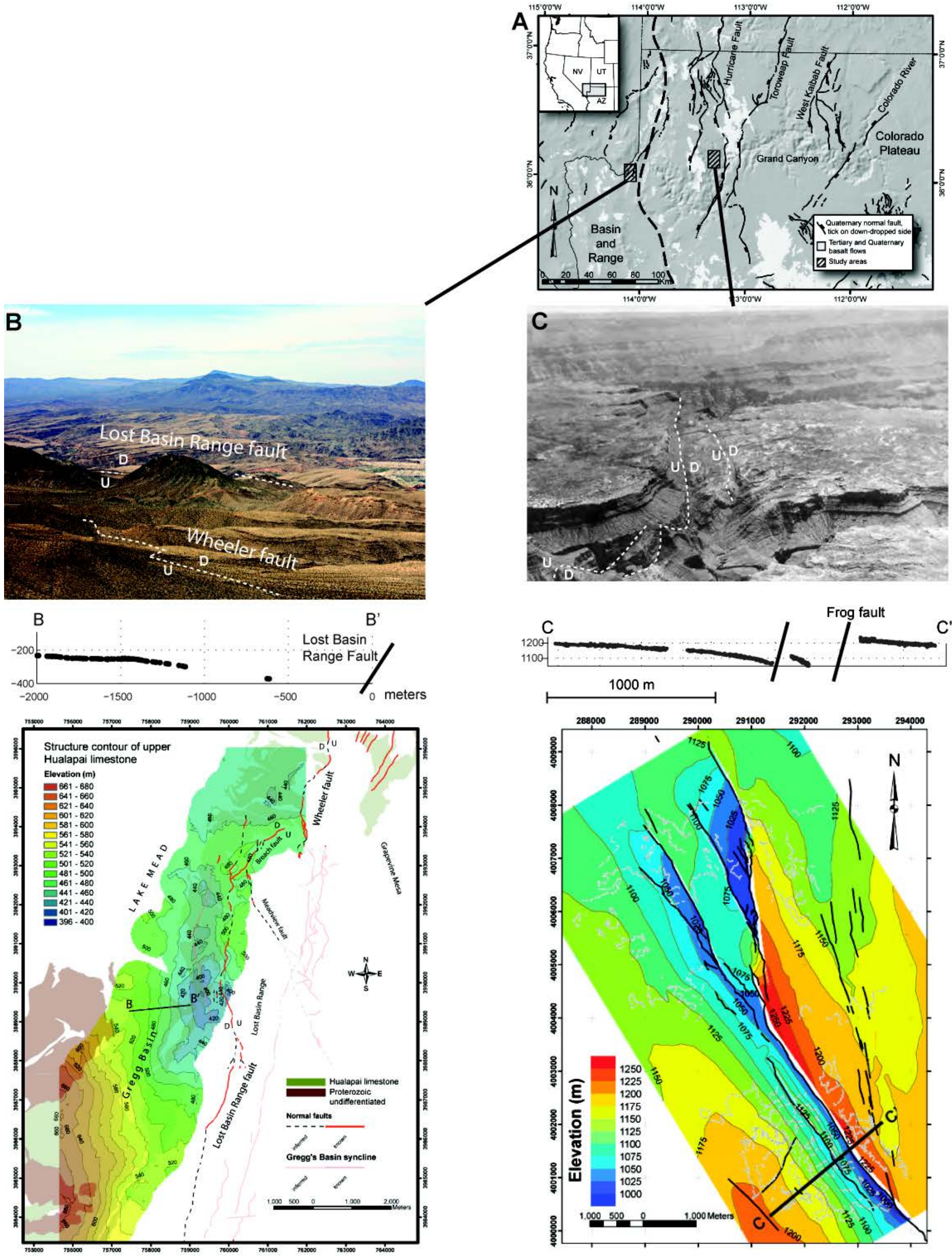

Figure 1. A. Study areas outlined on shaded relief map of the Grand Canyon and eastern Basin and Range (see Resor, 2008 for data sources). B. Folding of the Hualapai Limestone along the Wheeler fault system. Top: View to southwest of relay between Wheeler and Lost Basin Range faults from Meadview overlook. Middle: GPS Profile of upper Hualapai bed (looking north, see map below for location of profile). Bottom: Structure contour map of upper Hualapai Limestone derived from digital photogrammetry. C. Frog fault and Lone Mountain monocline. Top: oblique aerial photo (looking south, modified from Hamblin, 1965) of Frog fault (left), synthetic fault (right), and monocline. Middle: GPS profile of upper Esplanade bed (looking north, see map below for location). Structure contour of upper Esplanade Fm. derived from differential GPS surveying.

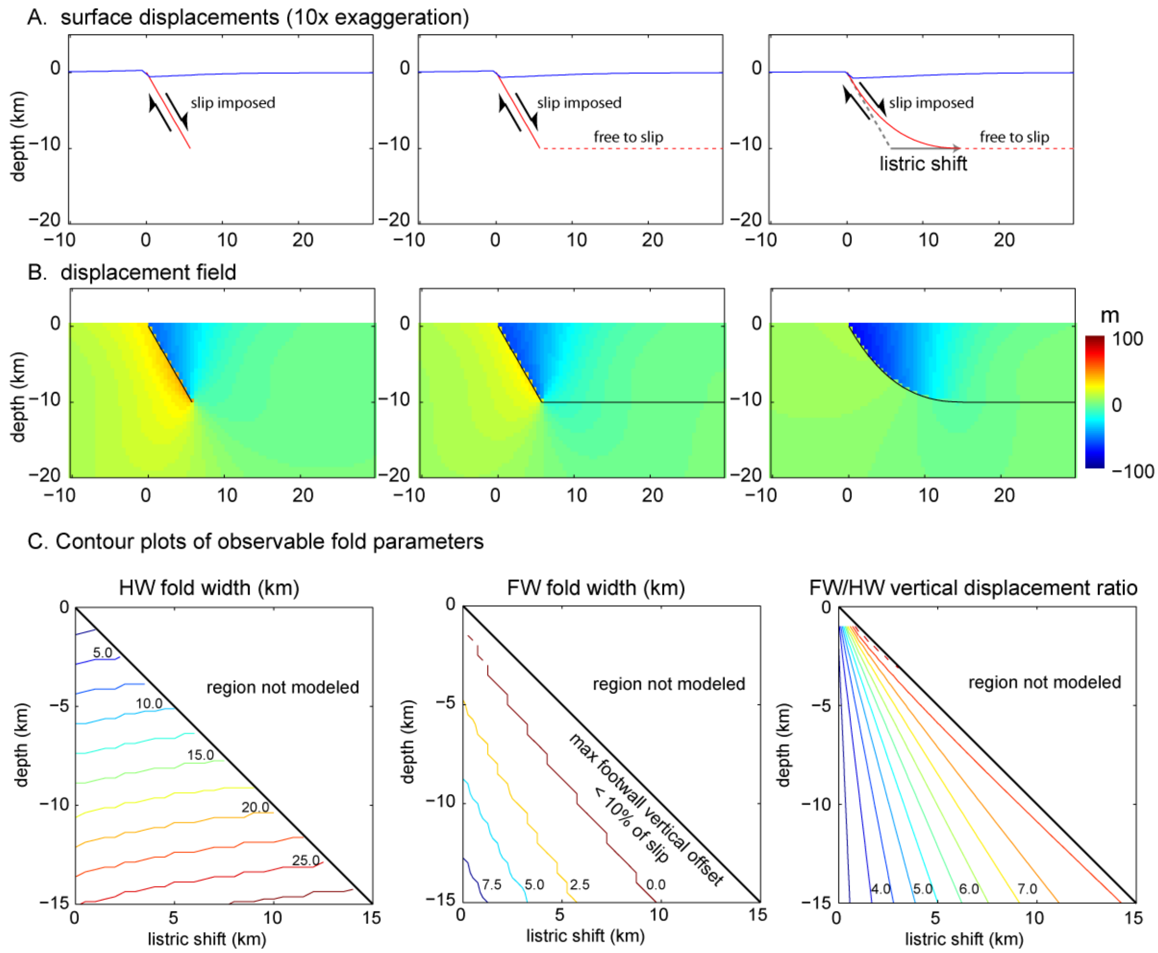

Figure 2. Elastic boundary element models of deformation associated with normal faults of varying down-dip geometry. A. Fault models (red) and ground surface with vertical displacement exaggerated 10x (blue). Left: finite planar fault model. Middle: finite planar fault ending at free-slipping detachment (red dashed line). Right: listric fault ending at planar detachment. Listric shift is defined as the horizontal offset between the lower tip of a planar fault and a listric fault of the same surface dip and depth. B. Vertical displacement fields for the models in A. C. Contours of fold parameters associated with varying listric shift and depth to detachment. Left: hanging wall (HW) fold width, measured as width from fault to point where vertical offset is less than 10% of throw. Middle: footwall (FW) fold width measured in a similar manner. Right: footwall/hanging wall vertical displacement ratio. Note that contour intervals vary between plots.

Cenozoic Evolution of the Grand Canyon and the Colorado Plateau Driven by Mantle Dynamics?

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A number of large depositional basins (e.g. Congo basin), large elevated orogenic plateaux (e.g. South African Plateau, Colorado Plateau), or deeply incised canyons (e.g. Grand Canyon) are difficult to explain by conventional tectonic mechanisms emphasizing forces associated with horizontal plate motions because such features are not situated on plates boundaries and evidence for crustal deformation of sufficient magnitude. Assuming that density heterogeneities in the mantle generate large-scale topography called dynamic topography [Conrad and Gurnis, 2003; Moucha and others, 2008; 2009], patterns of present-day deep asthenospheric flow are often proposed to explain such observed features [e.g Bird, 1984; England and others, 1998].

The Colorado Plateau area remains controversial and open to a wide range of plausible interpretations because its topographic evolution is linked with climate, erosion, near surface tectonics and mantle dynamics [e.g. McKee and McKee, 1972; Anders and others, 2005; Karlstrom and others, 2008]. Most of these studies try to link this evolution for the past 30 Ma with only the present-day mantle imagery, with poorly integrated dynamic considerations of the dynamic topography and mantle uplift through time. Despite numbers of studies [Sahagian and others, 2002; Libarkin and Chase, 2003; Sahagian and others, 2003; Flowers and others, 2008; Huntington and others, 2010], there is no consensus on the timing of Colorado Plateau uplift. However, three major stages have been presently proposed to explain the uplift of the Colorado Plateau to its current elevation of ~1.9 km: (1) an early Cenozoic (Laramide, 80-40 Ma) uplift related to Laramide low angle subduction [Bird, 1984; 1988] with crustal thickening [McQuarrie and Chase, 2000], or convective removal of lithospheric mantle [England and Houseman, 1988] or addition in the lithosphere of volatiles from the flat slab [Humphreys, 2003]; (2) 40-20 Ma uplift from buoyancy addition probably due to partial removal of the plateau lithosphere (Spencer, 1996); and (3) a late Cenozoic event related lithosphere heating from below [Thomson and Zoback, 1979] or convective removal of the lithospheric mantle with conductive mantle heating [Humphreys, 1995, Hinojosa and Mickus, 2002; Roy and others, 2009]. These three stages should be part of continuous evolution of the mantle convection.

Given recent advances in the tomographic interpretation of 3-D density variations in the mantle [Simmons and others, 2009], we now have the ability to simulate 3D mantle convection in response to the observed mantle density structure [Forte and others, 2007; Moucha and others, 2008; 2009]. By carrying out backward mantle flow simulations starting with present-day heterogeneity we can infer past mantle dynamics and related dynamic uplift/subsidence of the Earth's surface for the last 30 Ma [Forte and others, 2007]. However, testing such models, and especially mantle evolution through time, requires integrating detailed geologic data with model retrodictions (predictions backwards in time) of typically modest amounts of dynamic topography (hundreds of meters) through space and time [Moucha and others, 2009]. Moreover, the uplift history of the Colorado Plateau should strongly influence the evolution of the Grand Canyon. Thus, the evolution of the Colorado Plateau and the cutting of the Grand Canyon should be linked to the evolution of the dynamic topography in the past ~10-30 Myr.

Moucha and others [2009] reconstruct the motion of a warm mantle upwelling over the last 30 Ma toward the southwestern USA with a quantitative model of global mantle convection based on observed present-day tomography. Using these retrodictions, they compute the vertical stresses generated by viscous flow in the mantle in order to predict the amount of uplift/subsidence due to the mantle flow between 30 Ma and today. To be able to compare the dynamic uplift predictions to the field observations, the results are georeferenced through time by using the McQuarrie and Wernicke [2005] tectonic reconstruction of the southern USA. That may help reconcile diverse geologic, geomorphic and thermochronological datasets. This model of Colorado Plateau uplift (fig. 1) shows an average change in dynamic topography since 30 Ma on the order of 1000 m for the whole Colorado Plateau. However, the timing of predicted uplift varies between the different parts of the plateau [Moucha and others, 2009].

To illustrate the predicted uplift history for each model, we define three regions in the Colorado Plateau (Boxes N, SW, SE in fig. 2). In each region we extract the mean, minimum and maximum uplift in 5 Myr intervals. Figure 2 shows the evolution of the uplift since 30 Ma for each region. Although the total amount of uplift varies from about 700 m to 1200 m, the trend is similar in each model run: The uplift is relatively continuous since 30 Ma. However, little differences appear between the southwestern box and the northern box for the last 10 to 5 Ma: the southwestern box shows a decrease in uplift rate, and the northern box an increase of the uplift rate.

We also build a transect along the Colorado River from South West to North East and plot the uplift since 30 Ma along this transect for each 05 Ma time step (fig. 2). From 30 Ma to 15 Ma, the uplift is homogeneous all along the Colorado River transect. At 15 Ma, the uplift rate remains the same in the southern part of the transect (mouth of the present Grand Canyon) although it decreases in the northern part of the transect. Then, the uplift rate decreases a little in the southern part and increases incrementally to the northeast – the manifestation of a wave of uplift sweeping from southwest to northeast with a 200 to 400 m range. The northeastern propagation rate of the wave is around 20 km/Ma.

The retrodictions from the mantle convection modeling suggest a new interpretation of the Post-Laramide evolution of the Colorado River and SW Plateau that is consistent with essential geological constraints:

a) 30 Ma to 15 Ma: The dynamic uplift of the Colorado Plateau is regular and relatively homogeneous through the whole area, at a rate of 20-40 m/Ma, depending on model parameterization. The flow direction of the paleo Colorado River remains uncertain at this time [*Pederson*, 2008]. However, in details, the uplift is important in the eastern part of the Colorado Plateau. Thus, the paleo Colorado River does not probably drain the upper part of the Colorado Plateau, but only the area west of Kaibab monocline, as it has been pointed by Walcott and others [1890], Hill and others [2008; 2008a]. The erosion should be widespread through time as the uplift is continuous and mostly through the whole plateau. The paleo Colorado River could probably begin to slowly incise when the Grand Wash starts to be active, around 22 to 18 Ma [Lucchitta, 1972] and separates the Basin and Ranges province from the Colorado Plateau area.

b) 15 Ma to 10/5 Ma: At 15 Ma, the predicted uplift begins to be more important in the southwestern part of the Colorado Plateau than in the northeastern part, which causes the plateau to tilt drainages to incise eastwards. This may explain the Flowers and others [2008] AHe data and zonation (fig. 1): minimum partially reset ages are between 20 and 15 Ma old, and get younger to the northeast. During this uplift phase, the paleo Colorado River probably incises and builds a pre-Grand Canyon [Young and Brennan, 1974; McKee and others, 1967; Poliak and others, 2008; Flowers and others, 2008]. But this east-west differential uplift give the tilting that shuts off or strongly reduces the strength of the paleo Colorado River. The wave of uplift explains the incision of the early Grand Canyon and the reduced extension of the drainage area probably limited by the western part of the Kaibab monocline that remains a physical barrier for the eastern waters. It results in the formation of the lake Bidahochi in the east of Kaibab anticlines and thus the deposition of the Bidahochi formation, dated from 16 to 6 Ma [Dallegge and others, 2001]. The upper pre-Colorado River could flow eastwards.

c) 10/5 Ma to today: The wave of uplift moves from the west to the northeast at a rate ~10 km/Ma northeastwards. In consequence, the Colorado Plateau is tilted back to the west. That increases the incision rates of the Colorado River that forms the Grand Canyon. It also favors the capture of the upper Colorado River which drainage is reversed to the west. Due to age of the youngest lakes sediments, it seems that this wave begin to change the Colorado drainage system at about 8 Ma. The mechanism of the capture is not well constrained, but, according to Douglass [1999] and Meek and Douglass [2001], the lake Bidahochi spill over at around 5.5 Ma, that is coherent with dynamic uplift predictions. We hypothesize that this westward tilting and the capture of the Green River combine to trigger spill over of lake Bidahochi that leads to the dramatic post-6 Ma incision of the Grand Canyon (deepening the western Grand Canyon and forming the eastern Grand Canyon and Marble Canyon). In the global mantle convection models, the 5.5-5 Ma time correspond exactly to the most important change in dynamic uplift between the western part and the eastern part of the Colorado Plateau. This dynamic view of the evolution of the Colorado Plateau, predicts younging of the low-temperature thermochronological ages to the northeast, and is corroborated by the young AFT ages published by Stokli [2005] and Karlstrom and others. [2009].

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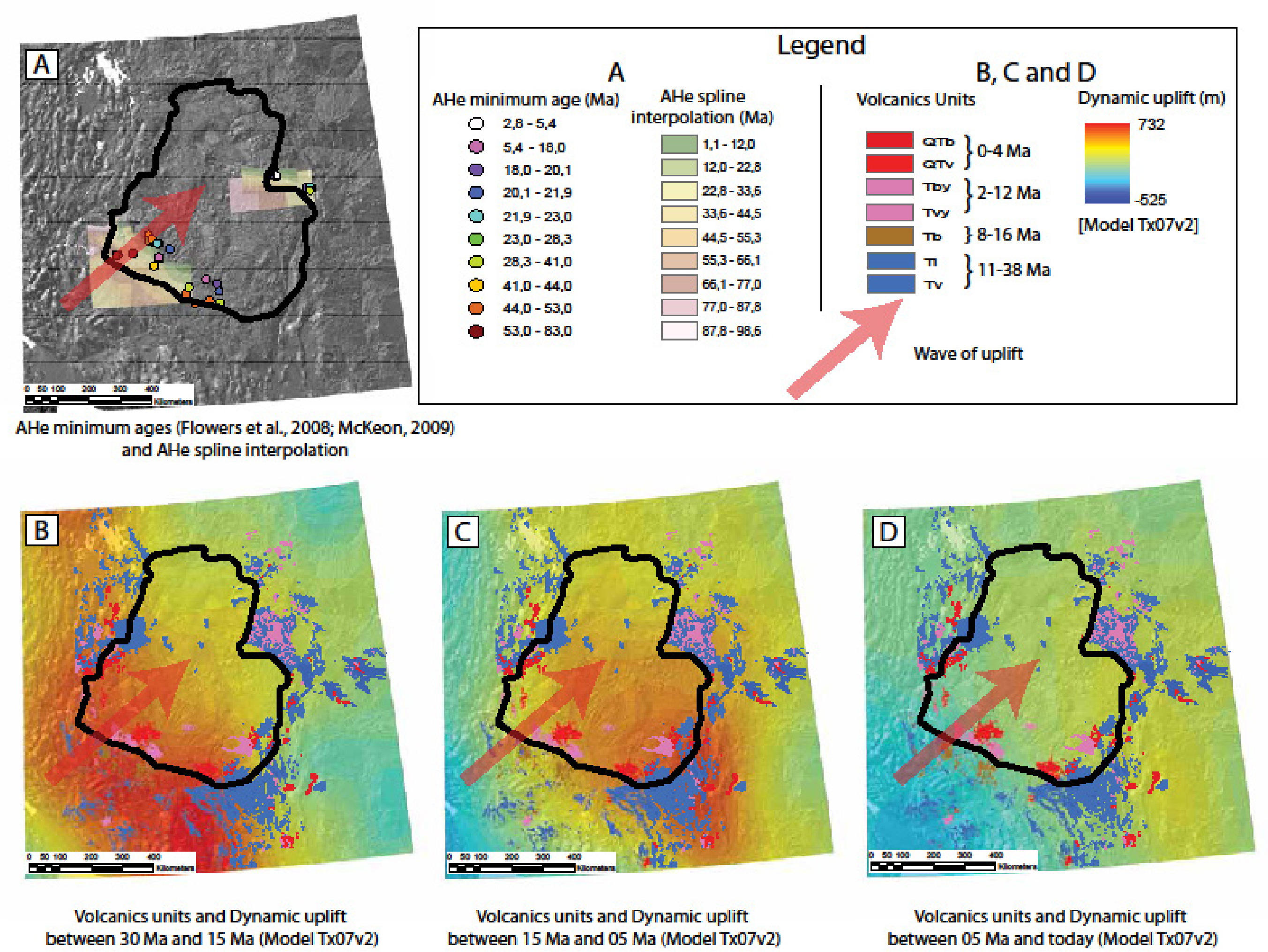

Figure 1. Evolution of the dynamic topography in the Colorado Plateau area. A) Apatite (U-Th)/He minimum thermochronological ages from Flowers and others [2008] and McKeon [2009] and tension spline interpolation using the 10 nearest points of the apatite (U-th)/He minimum ages. B) to D) Relationships between major volcanic units extracted from the USGS numerical geologic maps (Arizona - Colorado - New Mexico - Utah) and change in the predicted dynamic uplift for the model Tx07v2 respectively between 30 and 15 Ma, 15 and 05 Ma, 05 Ma and today. Each figure shows the shaded DEM in transparency, the contours of the Colorado Plateau and the direction of the wave of uplift.

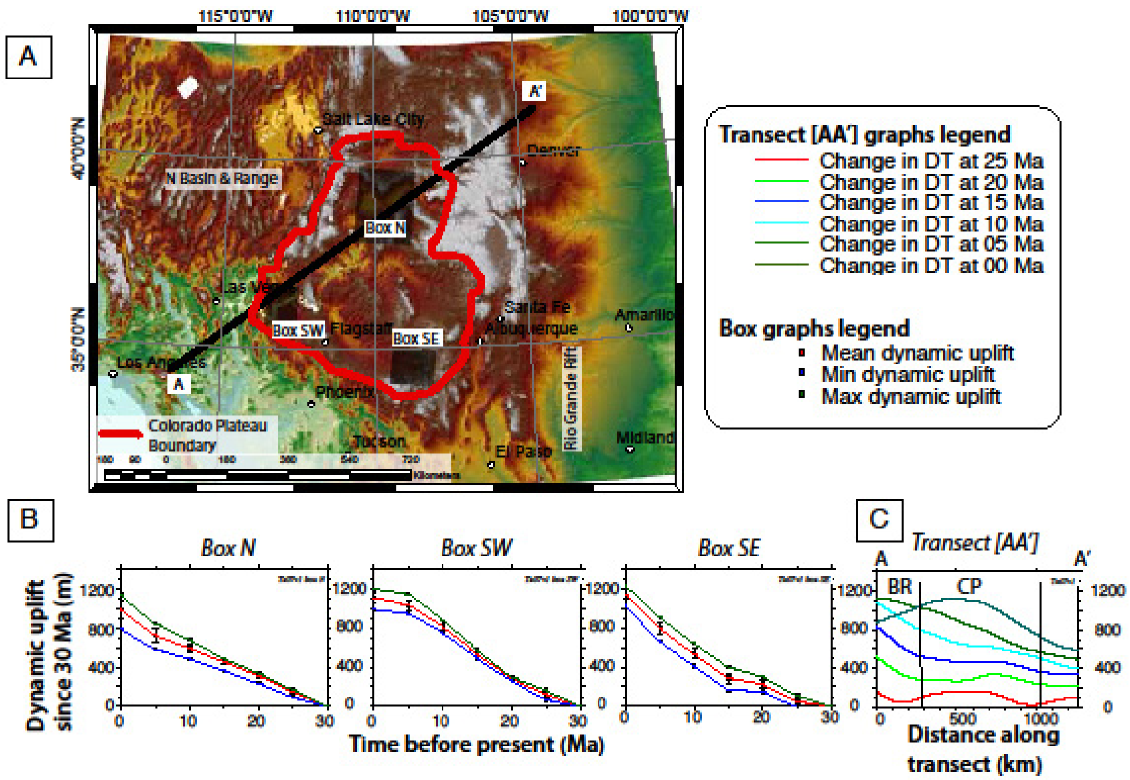

Figure 2. Dynamic uplift history since 30 Ma to today for the Colorado Plateau. A) Localization of the Colorado Plateau (red contour). The black boxes delimit the area where the uplift history is extracted for the four different models (in B). The black thick line tracks the transect [AA'] along which the amount of dynamic uplift since 30 Ma is extracted each 5 Ma and represented for the four models (in C). B) Uplift history since the last 30 Ma extracted from the 3 boxes showing the mean / maximum / minimum dynamic uplift respectively in red / blue / green. One sigma error bars are shown for the mean dynamic uplift. 0 Ma is today. C) Uplift history since 30 Ma along the profile [AA']. BR: Basin and Range; CP: Colorado Plateau.

Quaternary History of the Black Canyon of the Gunnison, Colorado

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The Gunnison River, a major tributary of the Colorado River flows through the Black Canyon, one of the narrowest (350 m) and deepest (700 m) bedrock canyons in North America. The modern longitudinal river profile (fig. 1) exhibits a prominent knickpoint within the Black Canyon. In the last 640 ka, average bedrock incision rates surrounding the knickpoint vary from 150 m/Ma (downstream, Darling and others, 2009), to 500 m/Ma (within, Sandoval, 2007), to 90-95 m/Ma (upstream, Aslan and others, 2008a). This pattern of fastest incision within the knickzone and slowest rates above necessitates upstream knickpoint propagation. A 640 ka terrace paleo-profile is reconstructed from numerous Gunnison River strath terraces containing Lava Creek B ash (Sandoval, 2007). This river profile reconstruction shows that a similar knickpoint existed at 640 ka in a location downstream from Black Canyon (fig. 1). Following abandonment of Unaweep Canyon at 1.06 ± 0.38 Ma (Aslan and others, 2008b) knickpoint propagation took place rapidly (>150 m/ka) in Cretaceous sediments, primarily Mancos Shale, and has been propagating more slowly since it encountered basement crystalline rock within the Black Canyon (fig. 1).

A constraint on the incision history of the Black Canyon comes from projecting the paleo Shinn-Boswick tributary to its intersection with the Gunnison River. This projection suggests 350-400 m of incision (of the ~700 m total depth) since abandonment of this paleotributary (fig. 2, Sandoval, 2007). We had previously reported an age of ~640 ka for the abandonment because Lava Creek B ash overlies (by 10 m) the Shinn-Bostwick paleo tributary to the Gunnsion River. The ash is within locally derived canyon fill that post-dated abandonment of the paleo-Bostwick River (Aslan and others, 2008). However, uncertainty about the age of the gravels and of abandonment has persisted because of the earlier report of 1.2 Ma Mesa Falls Ash a few meters below the Lava Creek B ash and above the Shinn-Bostwick gravels (Izett and Wilcox, 1982, based on Dickinson, 1966, *in* Hansen, 1967), which would yield slower incision rates.

New tephrochronology analysis of the original suspected Mesa Falls ash (sample obtained from Dickinson, 1966) showed that there is not a direct match between the Dickinson ash sample and the Mesa Falls ash in the USGS database. However, there were a number of good (>0.95, similarity coefficient) correlations to multiple Lava Creek ash bed samples in the database. The Dickinson sample exhibited a lower iron concentration level that is more similar to Mesa Falls glass shards than Lava Creek B, as well as the moderate hydration typically seen in older samples. Given that Yellowstone tephra (Huckleberry Ridge, Mesa Falls, and Lava Creek) have overlapping chemistries, the original identification as Mesa Falls ash was probably based on these latter characteristics. Although a Mesa Falls correlation remains possible from a tephrochronology viewpoint, new field mapping has failed to reveal a marked unconformity in the ~5-6 m of fill between the ashes, such that the correlation to Lava Creek B ash is supported by the stratigraphic context (Aslan and others, 2008a).

In addition, we have obtained a new cosmogenic burial age from a Bostwick Park gravel quarry that provides evidence against a possible Mesa Falls correlation for the Dickenson ash. Quartzite cobbles were sampled from the bottom of a 10 m thick gravel deposit in an active quarry. The gravels underlie Lava Creek B ash (640 ka, Lanphere and others, 2002; Sandoval, 2007). These gravels yield an isochron cosmogenic burial date of 870 ± 220 ka, within error of Lava Creek B, but not Mesa Falls ash. This age yields an incision rate of about 400 m/Ma since abandonment of the Bostwick paleotributary. Combined geologic and analytical uncertainties are fairly large such that possible incision rates are bracketed between 1000 m/Ma (350 m in 650 ka) to 367 m/Ma (400 m in 1090 ka). But, given rates of 400-500 m/Ma, if one assumed steady average rates over this time interval, this would indicate that Black Canyon has been carved in the last 1.37-1.71 Ma, in agreement with Hansen's (1967) estimate.

Downstream from the Black Canyon, ten strath terraces ascend from the North Fork of the Gunnison River-Gunnison River confluence to 670 m above the modern river. A cosmogenic burial date of 1 Ma on the seventh terrace anchors the 640 ka paleo-profile (mapped as Qt 5/6 in Sandoval, 2007) giving an average incision rate for the Gunnison River of 220 m/Ma. Approximately graded with the North Fork terrace, the Redlands Mesa pediment is tentatively assigned the ~1 Ma age. Strath terraces and pediments are inferred to record glacial-interglacial stages.

Longer-term incision rates on the Gunnison River also support models for knickpoint transience over the time span of 10 Ma. Rates from gravels below the 10 Ma Grand Mesa basalt are ~150 m/Ma (without an assumed depth to bedrock; Darling and others, 2009), whereas rates from gravels below the 10 Ma basalt flows on Flat Top and Red Mountain near the Gunnison River are ~55 m/Ma. An Oligocene paleo-Gunnison River was in approximately its present course ~30 Ma as indicated by ash flow units (Hansen, 1967), but the bedrock strath was ~500 m lower than at 10 Ma (fig. 1); hence bedrock incision rates were slow (or negative due to surface uplift) and river gradients were low from 30 Ma to 10 Ma. Although driving forces remain poorly constrained, we propose a model that involves the following interacting forcings: 1) knickpoint migration from 10 Ma to today, reflecting base level-fall due to epeirogenic uplift of the Rockies, 2) drainage reorganization of the Gunnison and Uncompagre River systems due to the abandonment of Unaweep Canyon ~ 1 Ma, 3) carving of much of the Black Canyon in the last 1.4 Ma , 4) elevation of Grand Mesa due to isostatic response to denudation, and 5) superposition of glacial-interglacial climatic cycles on the tectonically-driven landscape incision and denudation.

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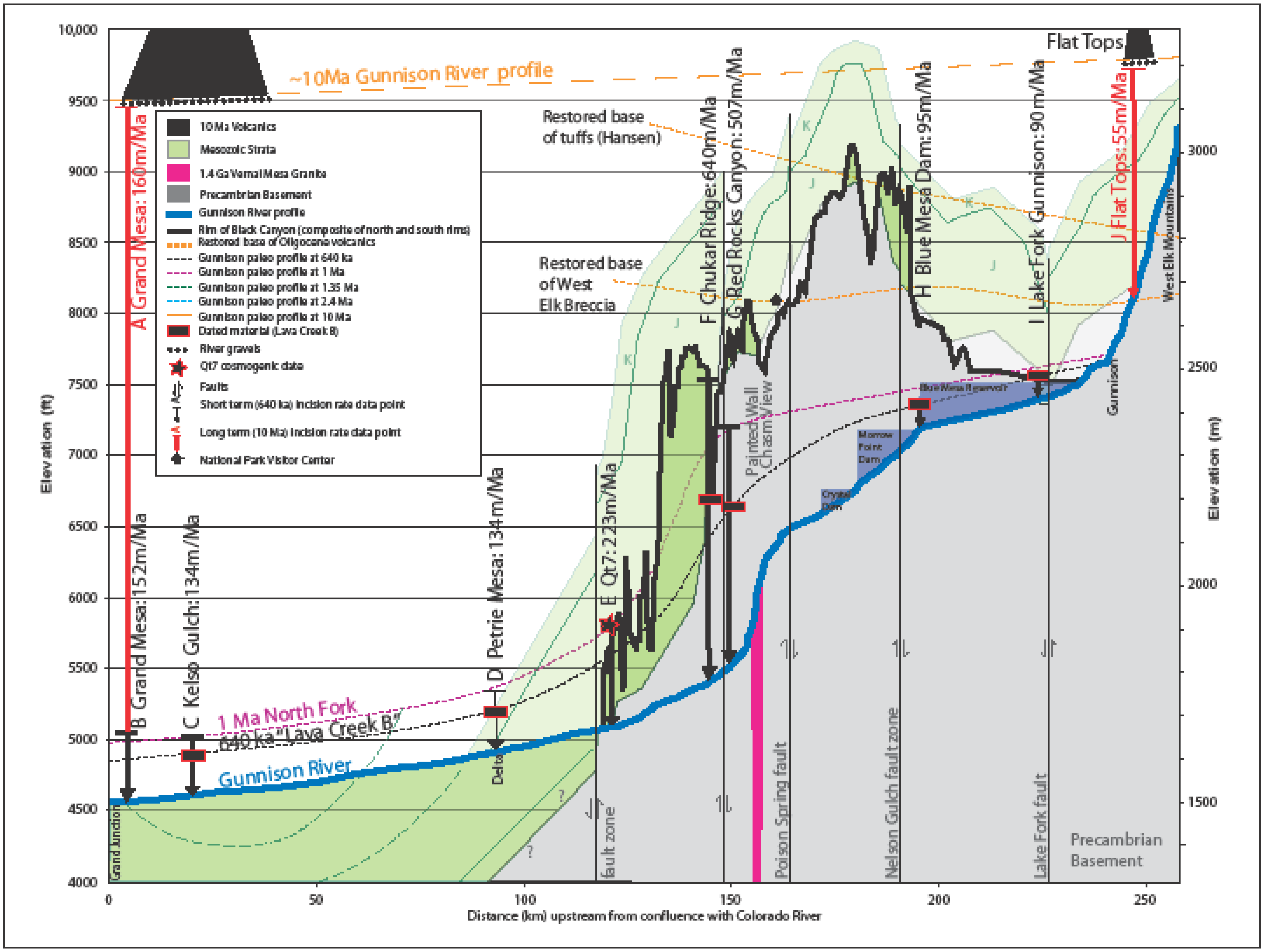

Figure 1. Longitudinal profile of the Gunnison River through Black Canyon; height of canyon walls and bedrock type are also shown. Incision rate vectors are shown for both 10 Ma timescales (red arrows) and 640 ka time scale (black arrows), These data allow the approximate reconstruction of the 640 ka paleo profile. Strath heights for 640 ka terraces are shown as black/red. Blue star is the position of the new 870 ka cosmogenic burial date (previously shown at 640 ka).

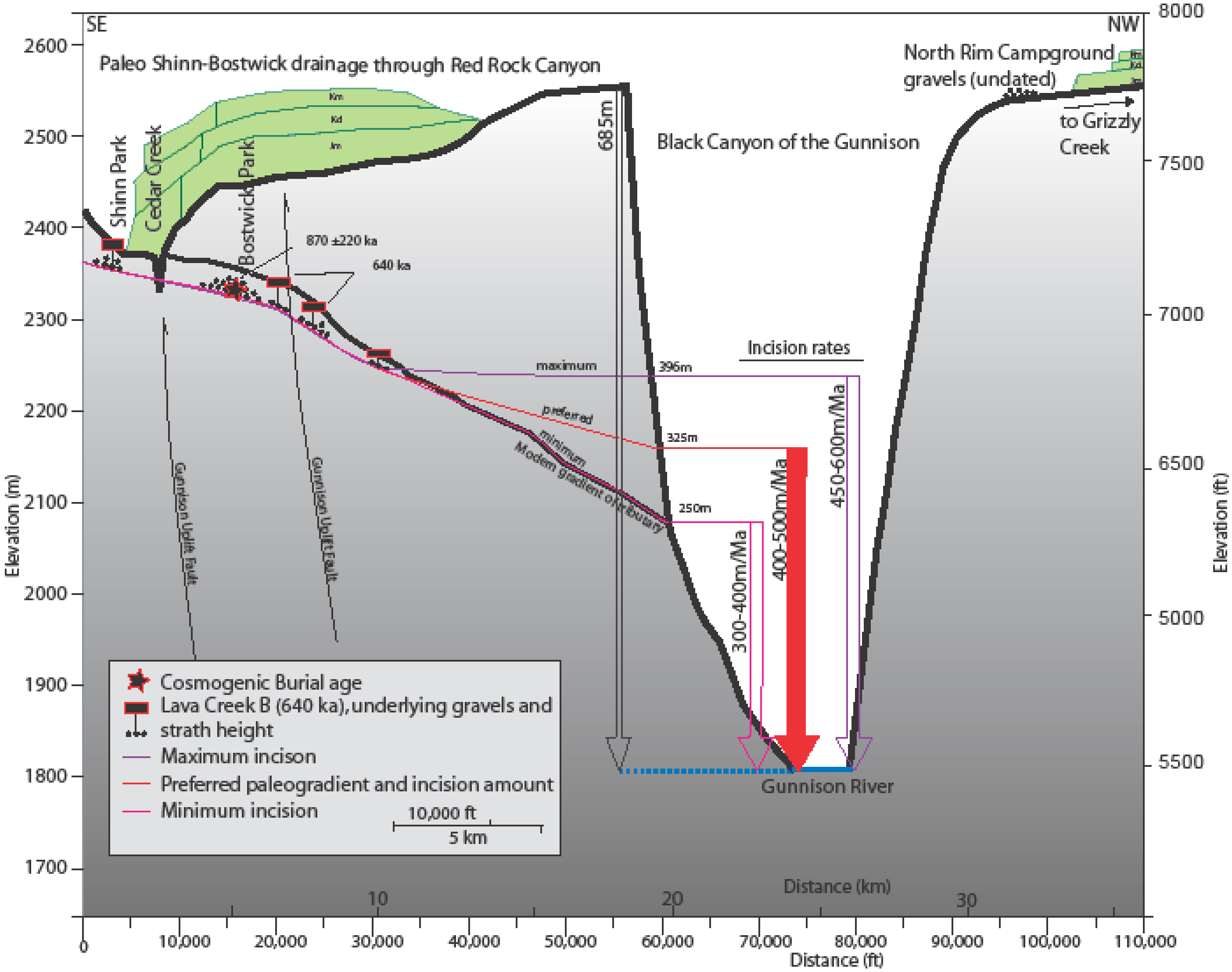

Figure 2. From Sandoval (2007) and Aslan and others (2008a) shows profile of paleo Shinn-Bostwick River confluence with the Gunnison River through Red Rock Canyon. Locations of Lava Creek B ash and cosmogenic dating are indicated. Minimum, maximum and preferred incision rates are shown.

A Brief Review of Sr Isotopic Evidence for the Setting and Evolution of the Miocene-Pliocene Hualapai-Bouse Lake System

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Introduction

Latest Miocene to earliest Pliocene inundation of the lower Colorado River trough and deposition of the Bouse Formation has been attributed to filling of closed basins by first-arriving Colorado River water, or to regional subsidence resulting in marine incursion during early opening of the Gulf of California. A lacustrine origin is supported by Sr, O, and C isotopic evidence (Spencer and Patchett, 1997; Poulson and John, 2003; Roskowski and others, 2010), consistent maximum elevations of Bouse deposits within proposed paleolake basins (Spencer and others, 2008), and sedimentological evidence of floodwater influx derived from northern sources immediately preceding Bouse deposition in Mohave and Cottonwood Valleys (House and others, 2008). A marine origin is supported by the presence of some marine species represented by fossils and shells from low elevations in the axis of Blythe basin, which is the southernmost of the Bouse basins (Smith, 1970; Todd, 1976; McDougall, 2008). In this extended abstract we briefly review Sr isotopic data that support a lacustrine origin for the Bouse Formation with sequential filling and spilling of a chain of lakes (fig. 1), and consider the role of the Hualapai Limestone in initial arrival of Colorado River water to the Mojave Desert region.

Hualapai Limestone

The Hualapai Limestone is exposed extensively near the mouth of Grand Canyon in Grand Wash trough and farther west in the central Lake Mead area (Beard and others, 2007; fig. 1). In Grand Wash trough, Hualapai Limestone was deposited from >11 Ma to <7.4 Ma and is exposed at altitudes ranging from ~530 to 912 m asl (Wallace and others, 2005). Hualapai Limestone is exposed in the central Lake Mead area at altitudes up to 720 m asl; deposition continued in this area until <6 Ma (Spencer and others, 2001). Lack of sediment derived from the Colorado River and facies relationships indicate that the Colorado River did not enter the Grand Wash trough during most or all of the time of Hualapai Limestone deposition (Lucchitta, 1987), although farther west rounded river gravel conformably overlies Hualapai Limestone deposits locally (Howard and Bohannon, 2001; Howard and others, 2008).

$^{87}Sr/^{86}Sr$ values of four Hualapai Limestone samples from Grand Wash trough decrease from 0.7195 at ~11 Ma to 0.7154 at ~7 Ma (fig. 2). Sr ratios were also determined for a section in the Temple Bar area of central Lake Mead where Hualapai Limestone outcrops range in altitude from about 550 to 715 m asl. A tephra collected from the middle of this section (660 m asl) has been dated at 6 Ma (Spencer and others, 2001). $^{87}Sr/^{86}Sr$ values from the Temple Bar section are less than those of Grand Wash trough and, with significant scatter, decrease upsection (fig. 2;

some overlap in $^{87}Sr/^{86}Sr$ values between Grand Wash Trough and central Lake Mead samples is recognized in new data – Karl Karlstrom, written communication, 2010). Based on currently published data, Hualapai $^{87}Sr/^{86}Sr$ values generally decreased with time and all values are substantially higher than Colorado River water (0.7103 to 0.7108) and Bouse carbonates (0.7102 to 0.7114). The lower maximum altitude (~720 m vs. 912 m) and lower Sr values of age-equivalent or younger Hualapai deposits farther west is compatible with a model whereby the Grand Wash paleolake spilled over to the west at ~7 Ma (possible paleodivide shown in figure 1), forming a separate, downstream lake (Lake Hualapai in fig. 1). This model presumes that down-to-west displacement associated with the Wheeler fault zone just west of Grand Wash trough is localized near the fault and has not resulted in substantial relative lowering of the entire central and western Lake Mead area.

Gradually decreasing $^{87}Sr/^{86}Sr$ values and a substantial increase in inundation extent may indicate increasing influx of water from a larger source region, possible including incipient Colorado River inflow, and mixing of this water with lake water derived from deeply circulating groundwater that had acquired high $^{87}Sr/^{86}Sr$ values from Proterozoic crystalline rocks (e.g., Crossey and others, 2006; Faulds and others, 2001, 2008). We have not found $^{87}Sr/^{86}Sr$ values that are equivalent to Colorado River values in the highest preserved Hualapai Limestone outcrops in the Temple Bar area. We speculate that the Colorado River abruptly entered the Hualapai lake system, leading to immediate lake overflow, incision of an outflow channel, draining of Lake Hualapai, and termination of lacustrine sedimentation.

Las Vegas area

At Frenchman Mountain east of Las Vegas (fig. 1), an algal limestone that is younger than 5.6 Ma (Castor and Faulds, 2001) and thus roughly equivalent in age to the Bouse Formation (House and others, 2008), yielded $^{87}Sr/^{86}Sr$ values of 0.7107 to 0.7109 (Roskowski and others, 2010). These ratios are well within the range of Bouse values and are similar to values for modern Colorado River water, but are lower than values obtained from the nearby Hualapai Limestone (fig. 3). Strontium isotopic data are thus consistent with a Colorado River source for the Frenchman Mountain algal limestone that reflects a new, post-Hualapai hydrologic regime. The fact that $^{87}Sr/^{86}Sr$ values are similar to Bouse values and unlike Hualapai values suggests that influx of voluminous Colorado River water overwhelmed the radiogenic Sr influx from the waters that had supplied Lake Hualapai. The one analysis of gypsum that underlies the Frenchman Mountain limestone (Duebendorfer and others, 2003) yielded slightly lower $^{87}Sr/^{86}Sr$ values than the values obtained from the limestone (fig. 3), further emphasizing the absence, or overwhelming dilution, of the radiogenic $^{87}Sr/^{86}Sr$ Hualapai water source during filling of Lake Las Vegas.

Blythe basin

Blythe basin is the most extensive basin containing Bouse strata, with abundant Bouse exposures at up to 330 m asl (fig. 1). Bristol basin is west of Blythe basin, but based on the altitudes of modern paleodivides and assuming no significant post-Bouse faulting or tilting, could have been the westernmost area of inundation during maximum filling of Blythe basin. A thin sequence of marl with basal tufa, located north of the town of Amboy in Bristol basin (fig. 1; marl initially discovered by David Miller, USGS), is lithologically similar to Bouse carbonates elsewhere. Isotopic analysis of five samples determined that $^{87}Sr/^{86}Sr$ values are similar to Bouse values

(one measurement) or somewhat elevated (four measurements; figs. 3, 4; Roskowski and others, 2010). A tephra within the marl contains glass that is geochemically correlated to glass within tephra found in the Bouse Formation at Buzzards Peak in the Chocolate Mountains (fig. 1; Spencer and others, 2001; Andre Sarna-Wojcicki and Elmira Wan, U.S. Geological Survey, written communication, 2008), and both have been geochemically correlated with the 4.83 Ma Lawlor tuff in the San Francisco Bay Area (Sarna-Wojcicki, 1976; McLaughlin and others, 2005; Andre Sarna-Wojcicki and Elmira Wan, U.S. Geological Survey, written communication, 2008). Lithologic similarity and geochemical correlation of the only known tephra in the Bouse Formation indicate simultaneous inundation of Blythe and Bristol basins and support the concept of extensive inundation of the lower Colorado River Valley and the eastern Mojave Desert during Bouse Formation deposition (fig. 1). Relatively high Sr isotopic ratios in Bristol basin may result from the influence of local radiogenic Sr sources within the somewhat isolated Bristol basin and poor mixing with Blythe basin water (Roskowski and others, 2010).
Two fossil fish (*Colpichthys regis*) from Bouse Formation marl near the south end of Blythe basin are interpreted as marine (Todd, 1976). Three samples derived from the two marl slabs that contain the fossil marine fish yielded $^{87}Sr/^{86}Sr$ values of 0.7107 to 0.7109 (Roskowski and others, 2010), similar to other Bouse Formation marl and to modern Colorado River water (Gross and others, 2001), but dissimilar to contemporaneous sea water (fig. 4). (The samples, collected in 1965 from the south end of Blythe basin, were provided by John Harris of the George C. Page Museum and Samuel McLeod of the Natural History Museum of Los Angeles County.) These new isotopic analyses support a saline lacustrine environment during Bouse deposition that was hospitable to a marine fish species.

An estuarine rather than lacustrine origin for the Bouse Formation was favored in Smith's (1970) paleontological study of the Bouse Formation partly because "The continuous occurrences of [the foram] *Ammonia beccarii* through thick sections of sediment indicate that conditions favorable to reproduction persisted for a long time." Such stable environmental conditions were considered unlikely in a lake. However, numerical simulation of arrival of Colorado River water and subsequent filling of previously closed basins along the Colorado River trough suggests that Lake Blythe could have had stable, approximately sea-water level salinities for tens of thousands of years (Spencer and others, 2008). Furthermore, *A. beccarii* can live and reproduce in both saline lakes (Cann and De Deckker, 1981) and at salinities less than 18 psu (practical salinity units – sea water is ~35 psu) (Takata and others, 2009).

Yuma area

Bouse Formation has been mapped adjacent to the Colorado River north of Yuma and south of the Chocolate Mountains paleodivide (Olmsted, 1972). Spencer and Pearthree recently examined these strata and concluded that the sediments are Colorado River channel sand and locally gypsiferous overbank mud. These strata grade upward into probable Quaternary tributary gravel deposits. Thus, these deposits do not represent the Bouse Formation and are probably substantially younger. All other inferred Bouse Formation strata in the Yuma area are in the subsurface (Olmsted and others, 1973; McDougall, 2008). We consider it a distinct possibility that lacustrine Bouse Formation strata are completely absent from the Yuma area, and that all marine shell fragments and microfossils in drill samples derived from subsurface units represent marine paleoenvironmental conditions near the north end of the early Gulf of California.

Blythe basin paleoenvironment

The presence of the planktic marine foram *Globigerina* sp., found primarily in drill samples from below sea level in the axis of Blythe basin, was interpreted by McDougall (2008) to indicate early marine inundation of the deep axis of Blythe basin, followed by influx of Colorado River water and development of less saline conditions. However, numerous Bouse carbonate samples from the lowest exposures in Blythe basin all contain $^{87}Sr/^{86}Sr$ values similar to Colorado River water and unlike seawater (fig. 4). Furthermore, early marine inundation followed by lacustrine inundation to 330 m altitude would require tectonic elevation of an edifice to impound Bouse lake waters after marine inundation and before Colorado River water arrived to fill Blythe basin. Construction of such a tectonic dam would presumably have resulted from fault-block uplift within the San Andreas transform fault zone, and likely would have required hundreds of thousands of years to develop. If this had happened, we would expect to find a lower unit of marine strata overlain by a higher unit of lacustrine strata, but even at the lowest exposures (70 m) in Blythe basin, Bouse carbonates resting on alluvial fan gravels have Colorado River type $^{87}Sr/^{86}Sr$ (Roskowski and others, 2010). We conclude that geologic and strontium-isotope evidence strongly support an entirely lacustrine origin for the Bouse Formation, and that paleoenvironmental conditions required for *Globigerinid* forams are less stringent than inferred by McDougall (2008).

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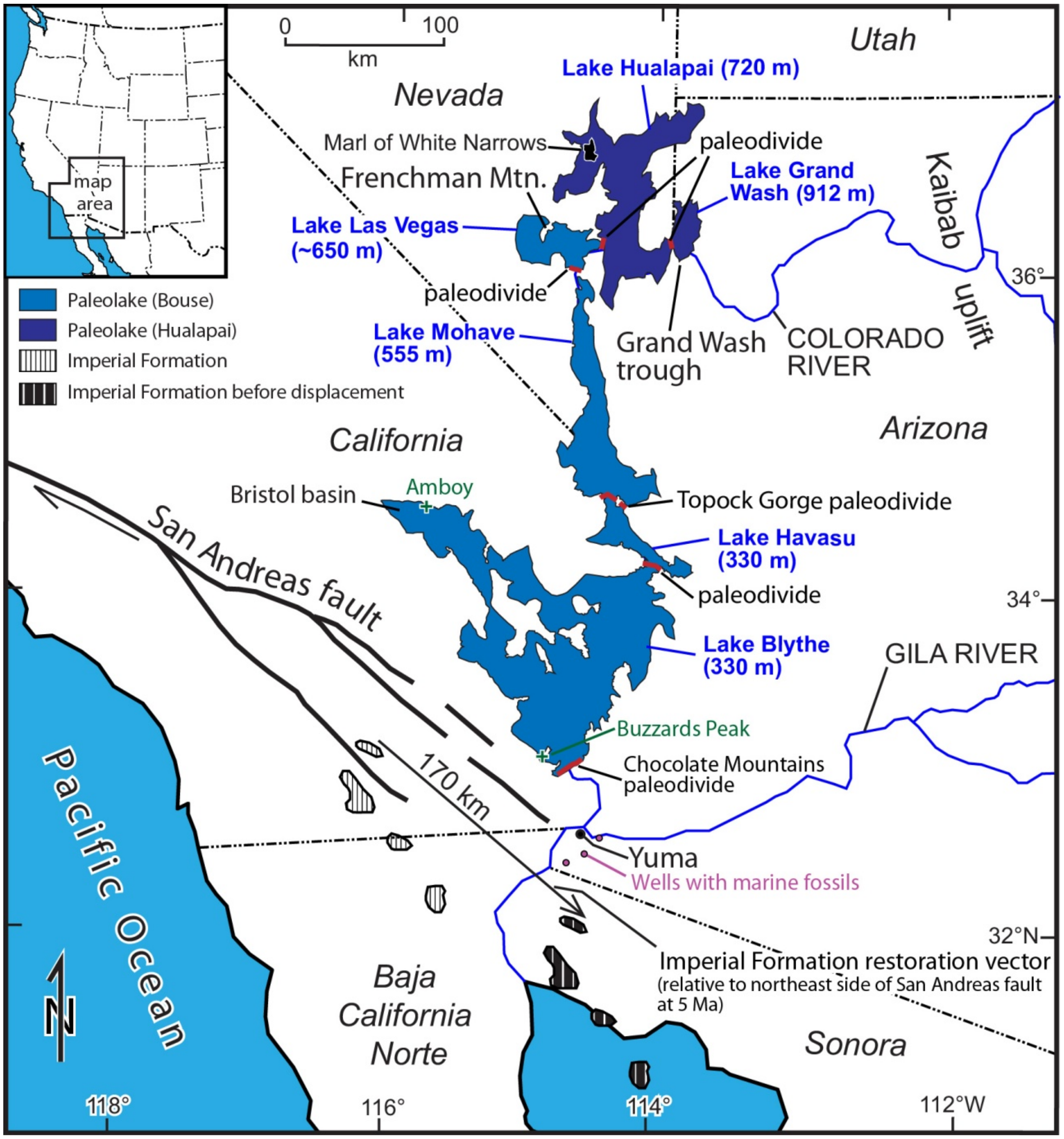

Figure 1. Hypothetical maximum extent of the Bouse-Hualapai lake system assuming modern elevations and no significant faulting-related elevation changes or tectonic tilting. All hypothetical lakes could not have been present simultaneously as evaporation would likely have been more than sufficient to remove all Colorado River water before spillover of Lake Mohave. Sequential filling and spilling of lakes in the Bouse-Hualapai lake system are inferred to have drained upstream lakes due to incision of outflow channels while filling downstream lakes. Evaporative concentration of salts during the fill and spill process resulted in near-marine salinity levels in Lake Blythe, which supported a small number of marine species likely introduced by birds (Spencer and Patchett, 1997; Spencer and others, 2008). Three wells in the Yuma area (purple dots on figure) recovered marine fossils interpreted to be older than the Bouse Formation farther north (McDougall, 2008).

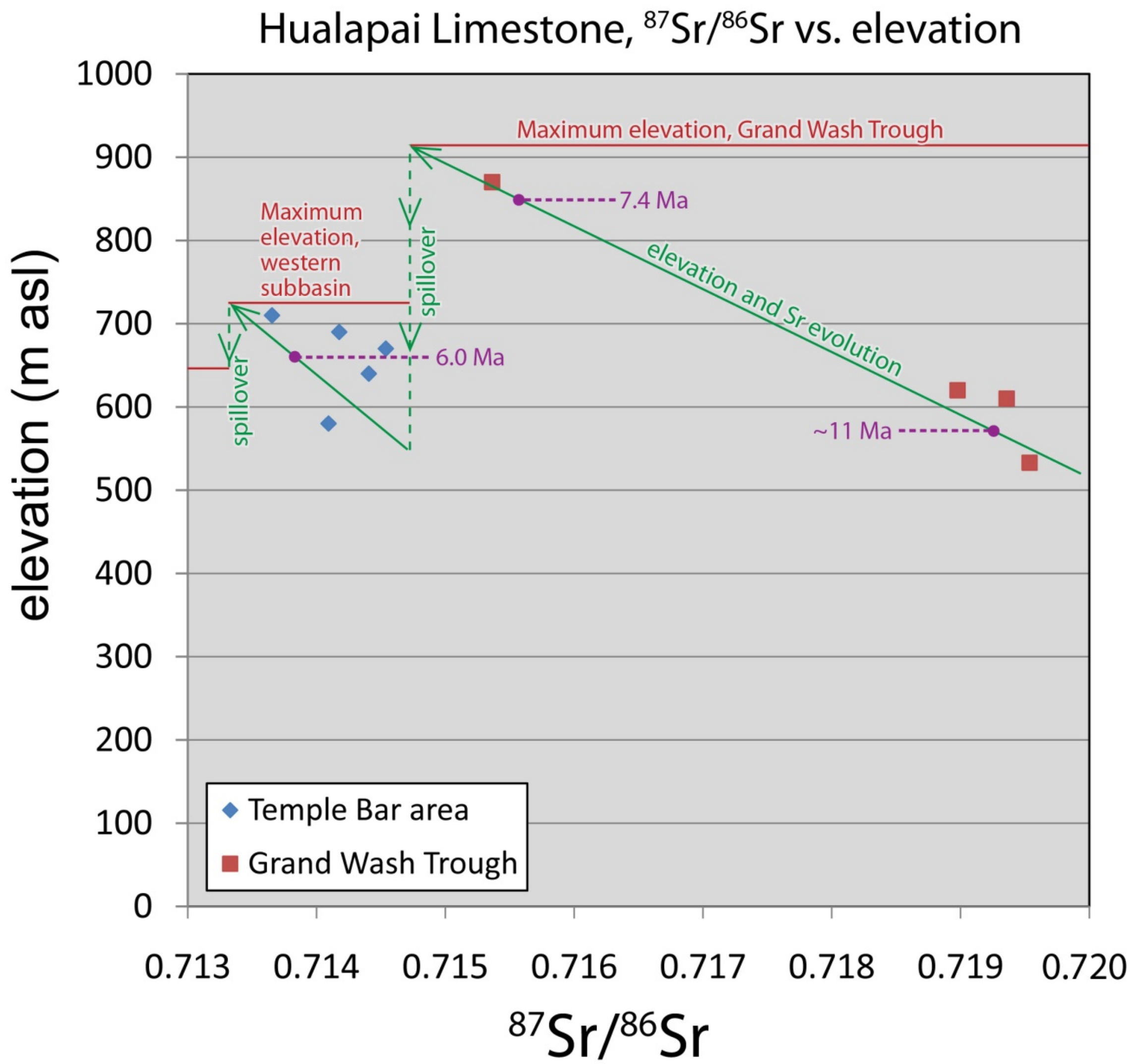

Figure 2. Postulated history of lake development and $^{87}Sr/^{86}Sr$ evolution of the Hualapai Limestone. Some scatter in points may result from the fact that elevation does not strictly reflect stratigraphic position. Data from Spencer and Patchett (1997) and Roskowski and others (2010).

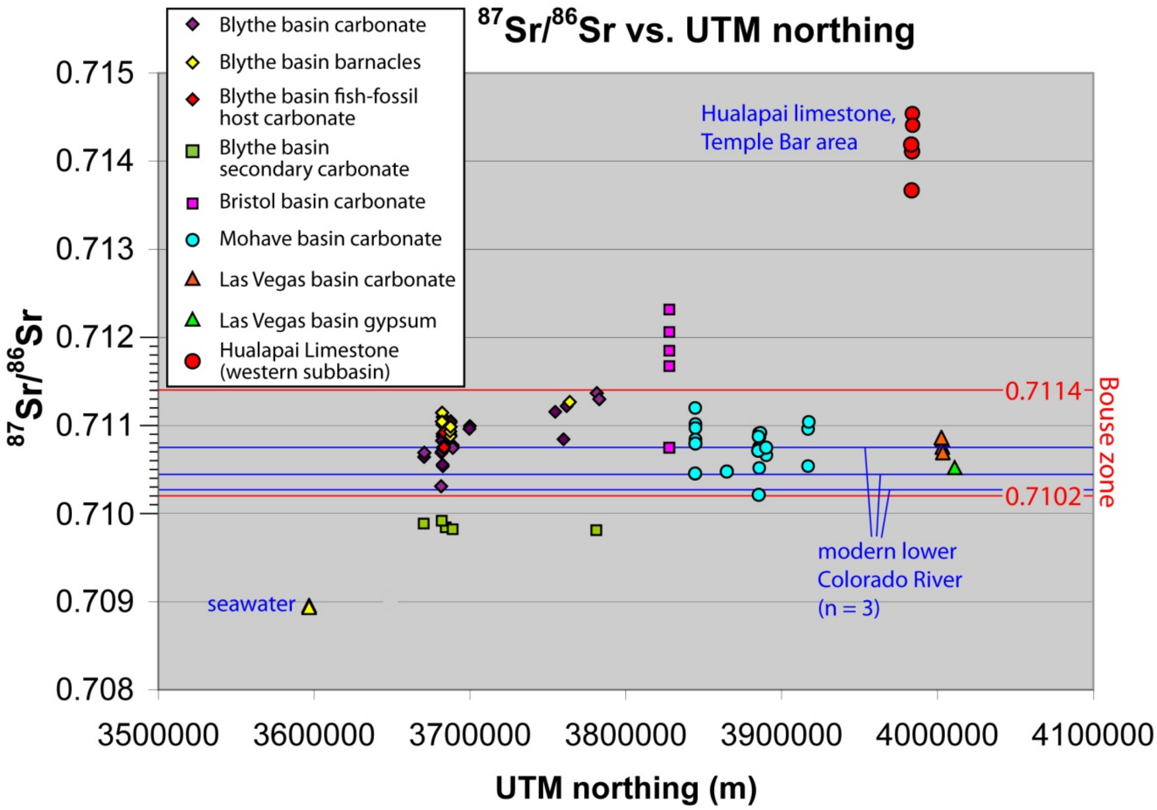

Figure 3. Sr isotopic composition of carbonate and shell samples vs. distance north (note that Hualapai Limestone samples are farther upstream along the Colorado River than the Las Vegas area samples, even though the Hualapai Limestone samples are from slightly farther south than the Las Vegas area samples). Analyses are from Buising (1988), Spencer and Patchett (1997), and Roskowski and others (2010). Also shown are values of modern lower Colorado River water (Goldstein and Jacobsen, 1987; Gross and others, 2001). Note the complete absence of any indication of reduced $^{87}Sr/^{86}Sr$ levels due to sea-water influx at southern locations. Similarly, $^{87}Sr/^{86}Sr$ at northern locations (Las Vegas basin) show no influence from water with elevated $^{87}Sr/^{86}Sr$ (Hualapai Limestone source waters) and are well within the expected range for Bouse lake waters and Colorado River water.

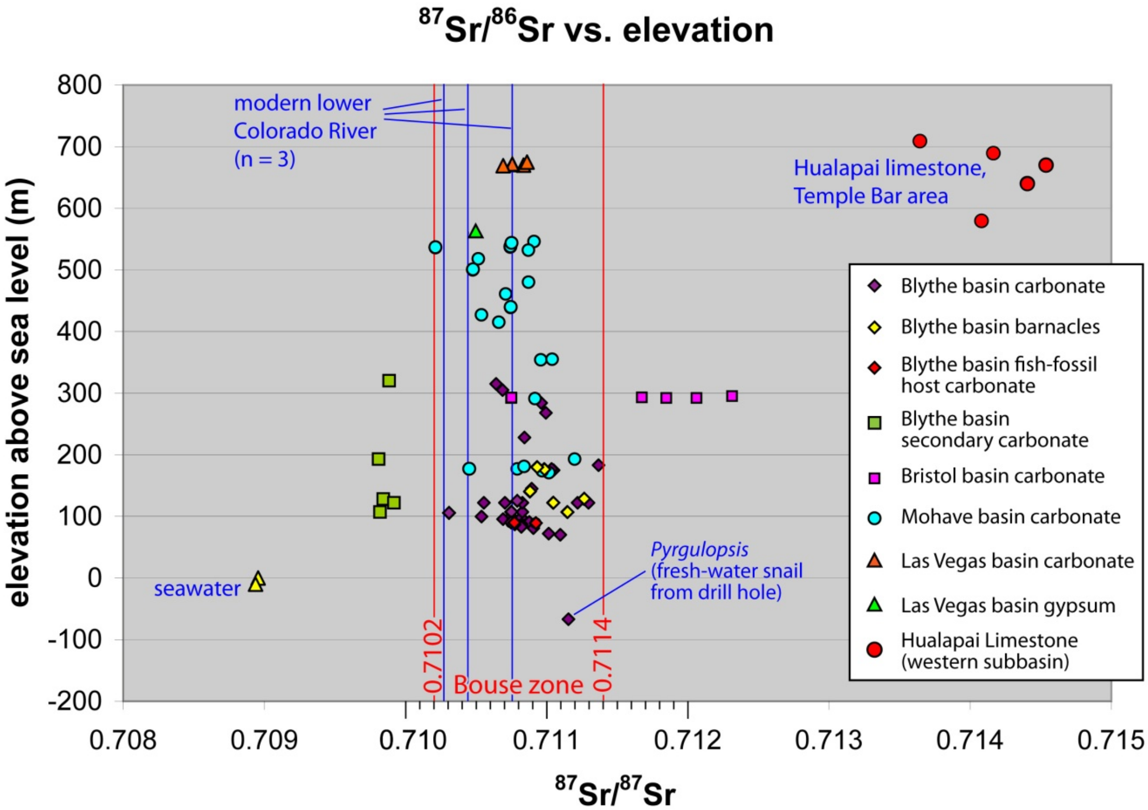

Figure 4. Sr isotopic composition of carbonate and shell samples vs. sample elevation. Analyses are from Buising (1988), Spencer and Patchett (1997), and Roskowski and others (2010). Also shown are values of modern lower Colorado River water (Goldstein and Jacobsen, 1987; Gross and others, 2001). Note the complete absence of any indication of reduced $^{87}Sr/^{86}Sr$ levels due to sea water influx at low elevations.

Updates on the Tectonics and Paleogeography of the Lake Mead Region from ~25 to ~8 Ma—Lakes and Local Drainages within an Extending Orogen, but no Through-going River?

Paul Umhoefer, Melissa Lamb, and L. Sue Beard

Editors of GSA Special Paper 463; "Miocene Tectonics of the Lake Mead Region, Central Basin and Range"

Many important refinements of past relations in the Lake Mead region and new models are presented in the GSA Special Paper 463 (Umhoefer, Lamb, Beard editors). The following conclusions are focused on aspects of the papers that relate to the paleogeography of the SW Colorado Plateau to central Basin and Range transition. Figures from the papers in the Special Paper that emphasize these points will be presented in a poster.

- Based on geophysics (Langenheim and others, 2010), the deepest basins near the SW edge of the Colorado Plateau that may have been drainage basins for pre Grand Canyon rivers are the Virgin depression (basin), northern Grand Wash, and Hualapai Basin (Red Lake).
- The oldest Cenozoic deposit in the Lake Mead region, the Rainbow Gardens Member of Horse Spring Formation was likely an internally drained basin in the topographic low north of the retreating Permian cliff of the Kingman high and east of the Sevier thrust front (Beard, 1996). The basin was a clastic to marshy setting from ~25 to 18.5 Ma; the basin was virtually all a carbonate marsh to lake from 18.5 to ~17 Ma (Lamb, Hickson and others unpublished).
- Reconstruction scenarios place the Rainbow Gardens basin in the Gold Butte to northern Grand Wash trough area (25 – 60 km north of western Grand Canyon) (Umhoefer and others, 2010), but the basin may have extended north to the Caliente volcanic field.
- Multiple datasets from thermochronology (Fitzgerald and others, 2009; Quigley and others, 2010; Karlstrom and others, 2010) and ongoing basin studies (Lamb and others, 2010; Faulds and others, 2010; Umhoefer and others unpublished) suggest that major extension started at ~17 Ma along the SW edge of the Colorado Plateau mainly on two fault systems (Grand Wash and South Virgin – White hills faults) that may have been linked to the north (Umhoefer and others, 2010).
- One model for the Proterozoic rocks of the Gold Butte block is that the South Virgin – White hills detachment fault projected up and flattened into a fault along the Great Unconformity (Karlstrom and others, 2010). An alternative model suggests that the South Virgin – White hills detachment fault connected above the Gold Butte footwall and into the subsurface to merge to the east with the Grand Wash fault as a classic core complex (Swaney and others, 2010).
- Another model suggests that detachment faulting dominated near the SW edge of the Colorado Plateau from 17 to 15 Ma, while transtensional faulting began at ~16 Ma and increased at 15-14 Ma as detachment faulting waned (Umhoefer and others, 2010). A fourth model links the South Virgin – White Hills detachment fault northward to extension in the Virgin depression via the Lake Mead fault system (Beard and others, 2010).
- Nothing from new data in the 17 – 14 Ma part of the Horse Spring Formation suggests the

presence of through–going drainages, but instead the basin(s) were dominated by lakes, marshes, and local streams in the west (Lamb and others, 2010; Hickson and others, 2010) and alluvial fans in the east (Blythe and others, 2010).

- The Bitter Ridge Limestone (~14.5 to ~13.5 Ma) was a lake in western Lake Mead that evolved from an open to closed basin due to faulting (Hickson and others, 2010). Continued faulting broke up the closed lake basin and formed complex facies in many environments from ~13.5 to ~10 Ma, but no evidence for a through-going drainage (Lamb, Hickson and others unpublished).
- From 14 to 12 Ma, eastern Lake Mead was dominated by large bajadas emanating from the footwall of the waning detachment and normal faults (Howard and others, 2010; Blythe and others, 2010).
- Voluminous volcanism across the southern Lake Mead domain (Smith and others, 2010; Faulds and others, 2010) from before 15 Ma to 10 Ma may have blocked through-going rivers from flowing toward the south.

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The California River and its Role in Carving Grand Canyon

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Recently published thermochronological and paleoelevation studies in the Grand Canyon region, combined with sedimentary provenance data in both the coastal and interior portions of the Cordillera, place numerous new constraints on the paleohydrological evolution of the southwestern United States. Review and synthesis of these data (Wernicke, 2011, *Geological Society of America Bulletin*, v. 123, doi: 10.1130/B30274.12011) suggest incision of a large canyon, from a plain of low elevation and relief to a canyon of roughly the length and depth of modern Grand Canyon, occurred mainly during Campanian time (80-70 Ma; figs. 1 and 2).

Incision was accomplished by a main-stem, northeast-flowing antecedent river with headwaters on the northeast slope of the Cordillera in California, referred to after its source region as the California River (figs. 1A and 2A, B). Thermochronological data indicate by the end of Campanian time (ca. 70 Ma), the river had cut to within a few hundred meters of its modern erosion level in western Grand Canyon near Precambrian basement. The, and In eastern Grand Canyon, the river had cut only down to lower Mesozoic strata, well above the modern erosion surface. Subsequent collapse of the headwaters region into a continental borderland, and coeval uplift of the Cordilleran foreland during the Laramide orogeny reversed the river's course by Paleogene time (figs. 1B and 2C).

After reversal, the terminus of the river lay near its former source regions in what is now the Western Transverse Ranges and Salinian terrane. Its headwaters lay in the ancient Mojave and Mogollon highland regions of Arizona and eastern California, apparently reaching as far northeast as the eastern Grand Canyon region. This system is also referred to after its source region as the Arizona River (figs. 1B and 2C, D). From Paleogene through late Miocene time, the interior of the Colorado Plateau was separated from the Arizona River drainage by an asymmetrical divide in the Lees Ferry-Glen Canyon area, with a steep southwest flank and gently sloping northeast flank that drained into large interior lakes, fed primarily by sources in the Cordilleran and southern Rocky Mountains to the north and west, and by recycled California River detritus shed from Laramide uplifts on the plateau (fig. 1B). By Oligocene time, the lakes had largely dried up and were replaced by ergs.

By mid-Miocene time, a pulse of unroofing had lowered the erosion level of eastern Grand Canyon to within a few hundred meters of its present level, and the Arizona River drainage system below modern Grand Canyon was deranged by extensional tectonism, cutting off the supply of interior detritus to the coast (fig. 2E).

Increasing precipitation in the Rocky Mountains in late Miocene time reinvigorated fluvial-lacustrine aggradation northeast of the asymmetrical divide, which was finally overtopped between 6 and 5 Ma, lowering base level in the interior of the plateau by 1500 m (fig. 2F). This event reintegrated the former Arizona drainage system through a cascade of spillover events through Basin and Range valleys, for the first time connecting sediment sources in Colorado with the coastal California. Combined with the intensification of summer rainfall as the Gulf of California opened, this event increased sediment flux through Grand Canyon by perhaps two orders of magnitude from its Miocene nadir, giving birth to the modern subcontinental-scale Colorado River drainage system. Thus, whereas the Colorado River system played a major role in unroofing the interior of the Colorado Plateau, it did not play a significant role in the excavation of Grand Canyon (figs. 2F, G).

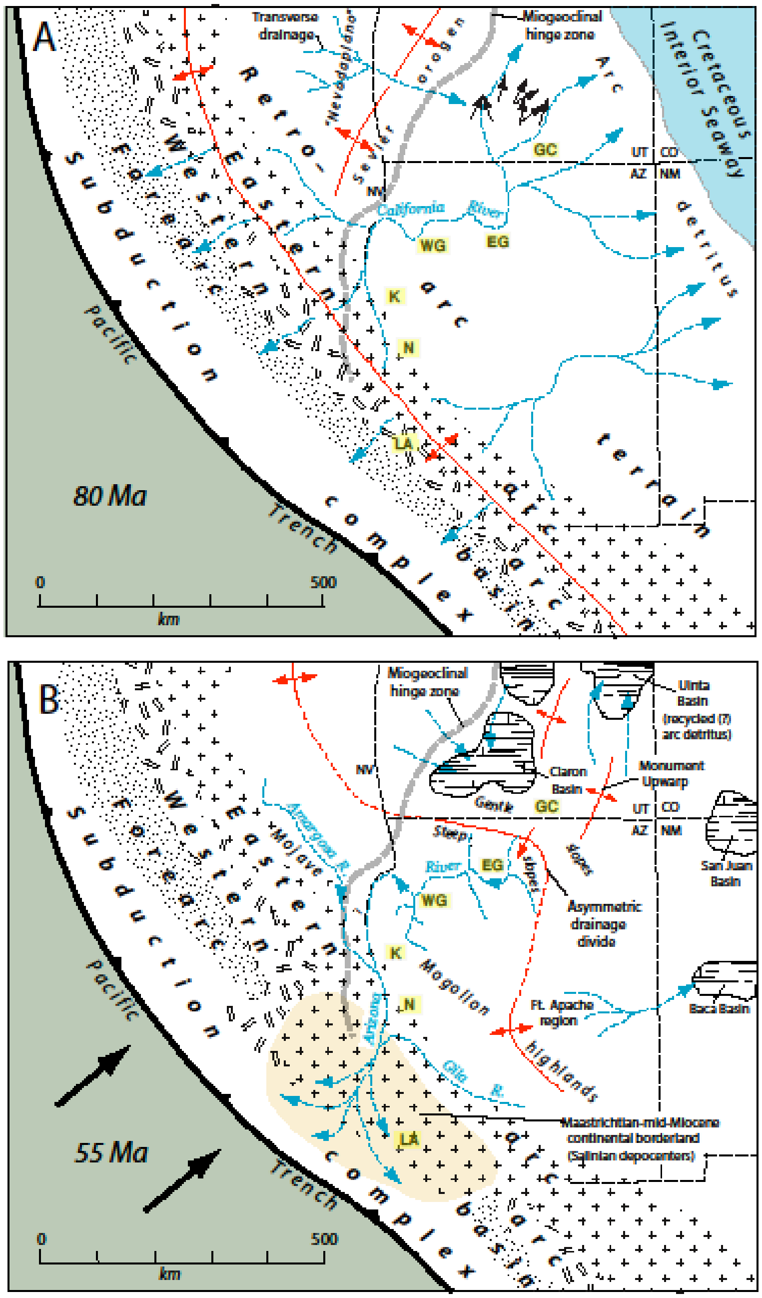

Figure 1. Paleogeographic maps showing: A) Campanian position of California River drainage near onset of incision in the Grand Canyon region. Red line shows position of drainage divide; black arrows show paleocurrent directions of the Wahweap Sandstone along with hypothetical position of drainage transverse to the Sevier orogen. B) Early Eocene positions of Arizona River drainage, the drainage system in the Fort Apache region and ancestral Gila and Amargosa rivers. Abbreviations highlighted with yellow background correspond to locations in figure 2 as follows: LA, Los Angeles; N, Needles; K, Kingman-Lake Mead area; WG, western Grand Canyon; EG, eastern Grand Canyon; GC, Glen Canyon. From Wernicke, B., 2011, Geol. Soc. Am. Bull., v. 123, doi: 10.1130/B30274.1 (2011).

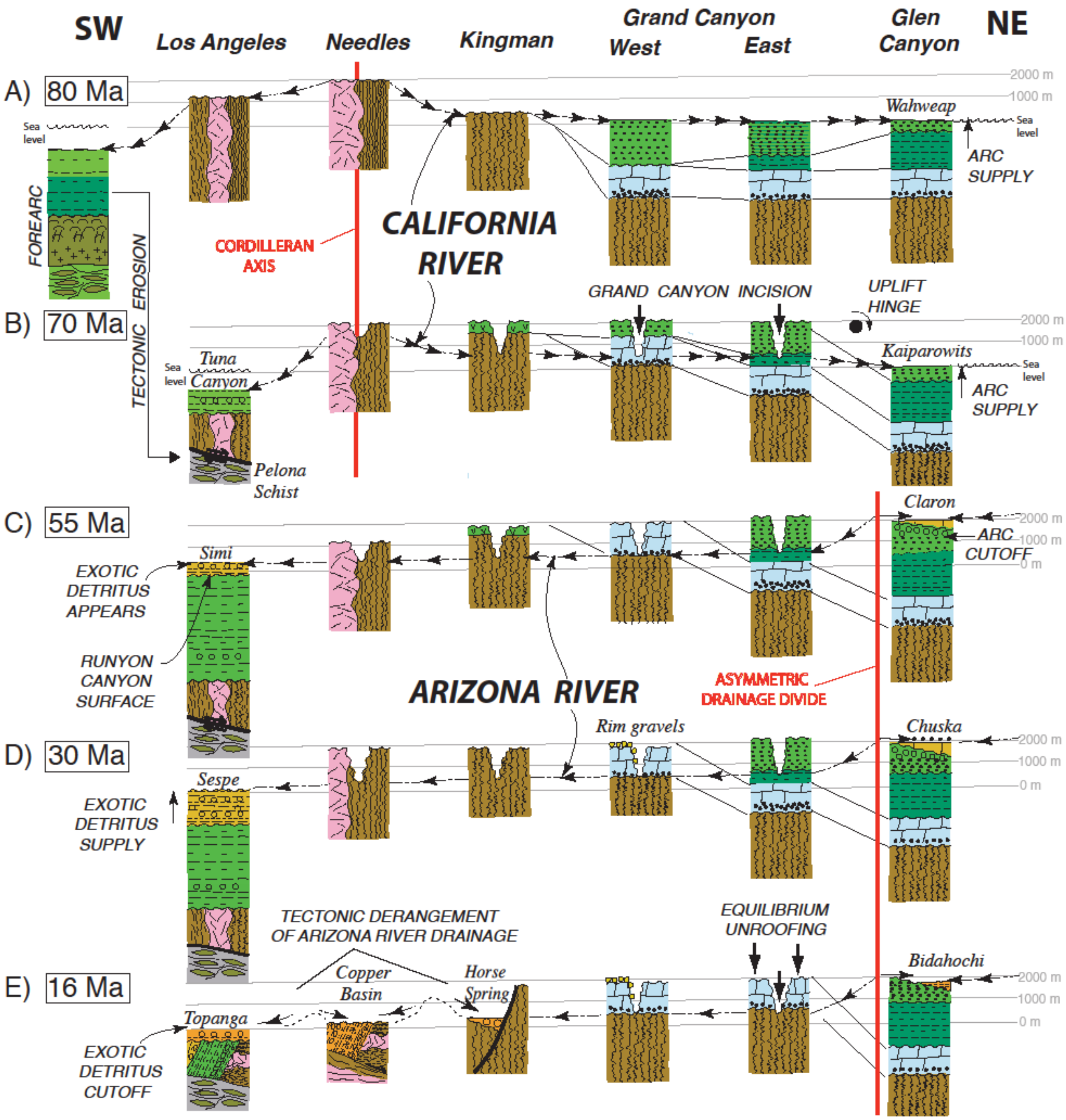

Figure 2. Diagrammatic cross sections of the six areas highlighted in yellow in figure 1. Wavy lines at the top of each diagram indicate the elevations of river grades and surrounding uplands, with key formations in depositional basins labeled in italics. Geologic units: Proterozoic (brown), Paleozoic (light blue), Triassic through Lower Cretaceous (forest green), Upper Cretaceous (chartreuse), Paleogene (gold), upper Oligocene through mid-Miocene strata (orange) and mid- to late Miocene strata (yellow), forearc ophiolitic basement (olive green), metamorphosed subduction complex (gray and olive), Mesozoic arc intrusives rocks (pink). Sediment load in rivers indicated qualitatively as high (three arrows), moderate (two arrows) and low (one arrow). Horizontal gray lines show elevation above sea level. From Wernicke, B., 2011, Geol. Soc. Am. Bull., v. 123, doi: 10.1130/B30274.1 (2011).

Oligocene Tuff Corroborates Older Paleocene-Eocene Age of Hualapai Plateau Basal Tertiary Section

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An Oligocene tuff collected in upper Peach Springs Wash in 2009 by R.A.Young and analyzed by L. Peters and R. Crow at the New Mexico Geochronology Research Laboratory provides the oldest $^{40}Ar/^{39}Ar$ age collected on any volcanic unit from the Hualapai Plateau in Arizona (figs. 1, 2, and 3) (Young, 1966, 2001). The tuff (figs. 4A, B, and C) is intercalated near the top of the thick sediments of the Peach Springs Member of the Buck and Doe Conglomerate, which overlies the conspicuous Eocene-Oligocene(?) disconformity (fig. 4D) that is approximately 90m below the ash in the Peach Springs section (Young, 1999; Young and Hartman, this volume). The age of the tuff was determined by single crystal laser fusion of 15 sanidine grains, yielding a weighted mean age of 23.97 ± 0.03 Ma (fig. 3). The newly discovered tuff is located in a section that also contains a basalt dated as 19.94 ± 0.4 Ma and that was incorrectly described as "overlying the Peach Springs Tuff" (Damon and others, 1996). However, as shown in figure 2, the Peach Spring Tuff (18.5 Ma) is not actually in direct contact with the local basalt flows. A slump block (fig. 2) containing the tuff occurs along the only access road down Peach Springs Wash and, as viewed from below, can leave an observer with the impression that local basalts actually do overlie the tuff. However, stratigraphic relationships there and further west along the Hurricane fault (fig. 2) show that the tuff is actually younger than the adjacent basalt outcrops. This clarification eliminates the discrepancy, or apparent age inversion of the tuff and basalts, as reported in Damon and others (1996).

The Peach Springs Member of the Buck and Doe Conglomerate records the local reemergence of post-Eocene (post-disconformity), northeast-flowing, local drainage into Peach Springs Wash from the adjacent Truxton Valley, immediately to the southwest (fig. 1). The Truxton Valley is partially ringed by outcrops of Precambrian rocks, including an unusual "stretch pebble" conglomerate from a locality locally known as "Slate Mountain" along the inferred basement extension of the Hurricane fault zone (Valentine SE Quadrangle mapped by Beard and Lucchitta, 1993). The Precambrian exotic clasts make the Peach Springs Member of the Buck and Doe Conglomerate appear superficially similar to the underlying, but much older, arkosic Music Mountain Formation, previously informally included in the Arizona "Rim gravels" (Young and Hartman, this volume). However, close examination of the complete Peach Springs Wash section (fig.2) demonstrates they are two vertically separate units, with different source rocks, and entirely different sedimentary characteristics (Young, 1966; 1999). The far-traveled, deeply weathered, igneous clasts of the older Music Mountain Formation crumble completely when collected (in situ) due to their extreme degree of weathering and much greater age, whereas the younger Buck and Doe clasts are not noticeably more weathered than similar crystalline cobbles that occur in recent alluvial deposits. This suggests a much greater antiquity (Paleocene-Eocene) for the basal arkosic conglomerates in the Tertiary section. The only well established age for the

Music Mountain Formation and proposed equivalents in northwestern Arizona is at Duff Brown Tank near Long Point, Arizona (Young and Hartman, this volume).

The position of the newly dated 23.97 Ma tuff so high in the local stratigraphic section (fig. 2) gives further credence to the argument that the reddish-hued exotic sediments below the regional disconformity are significantly older than Oligocene, and indicative of a very different climatic regime (Young, 1999; Young and Hartman, this volume). The stark contrast at the disconformity between the Milkweed Member of the Buck and Doe Conglomerate and the redder colors of both the underlying West Water Formation (paleosol capping Music Mountain Formation) and the contemporaneous Hindu Fanglomerate (local Paleozoic limestone clasts, fig. 2) is clearly illustrated in figure 4D. Given the relatively thin interval of sediments separating the Peach Spring Tuff (revised tuff nomenclature, Young 1999) from the newly discovered Oligocene tuff, and the intense weathering that characterizes the lower Paleogene section, it now is confirmed that the base of the Hualapai Tertiary section is significantly older (Paleocene-Eocene), as originally speculated by Young (1999, 2001) and as recently confirmed by Long Point limestone fossil ages from specimens collected at Duff Brown Tank (Young and Hartman, this volume).

The source of an Oligocene tuff in this location is unknown, but may not be far away. The tuffs of the Aquarius Mountains caldera (Young and McKee, 1978; fig. 3), partially mapped by G. Fuis (1974), are near the edge of the plateau approximately 28 miles (45 km) south southwest of Peach Springs, immediately south of Interstate Route I-40 (fig. 1). A thick sequence of rhyodacitic tuffs underlies the Peach Spring Tuff, which lapped onto the northern and western flanks of the Aquarius Mountains volcanic complex immediately following the main explosive phase of Aquarius Mountains tuff deposition. Nearby basalts, which rest on Precambrian basement rocks at the west edge of the Aquarius Mountains along the plateau boundary fault, have ages of 24.7 ± 3.5 Ma and 24.9 ±0.9 Ma (Young and McKee, 1978; Young, 2001, Appendix A, p. 244). (Bull Spring & Penitentiary Mt. Topographic Quadrangles, reconnaissance geologic maps, R.A. Young, 1978, unpublished).

The new Oligocene ash age significantly improves the chronology for the Hualapai Plateau Tertiary history. It provides a long-sought age for the Buck and Doe Conglomerate, which was previously only known to be younger than the inferred Paleocene or Eocene Music Mountain Formation and older than the Miocene Peach Spring Tuff. The new ash age, combined with the Eocene fossil ages (Young and Hartman, this volume), demonstrate that the Hualapai Plateau Tertiary sections preserve a relatively complete record from early Eocene or Paleocene time through late Miocene or early Pliocene time. The only indication of a significant hiatus in the Tertiary record preserved in the Laramide paleocanyons is the obvious disconformity (fig. 4D), which appears to be due to weathering and nondeposition rather than erosion in these isolated local basins (Young, 1999). The Hualapai Plateau sections in the Milkweed and Peach Springs paleocanyons appear to contain the most complete and best exposed Cenozoic record (potentially late Paleocene to Pliocene time) in northern Arizona, if not in the entire state.

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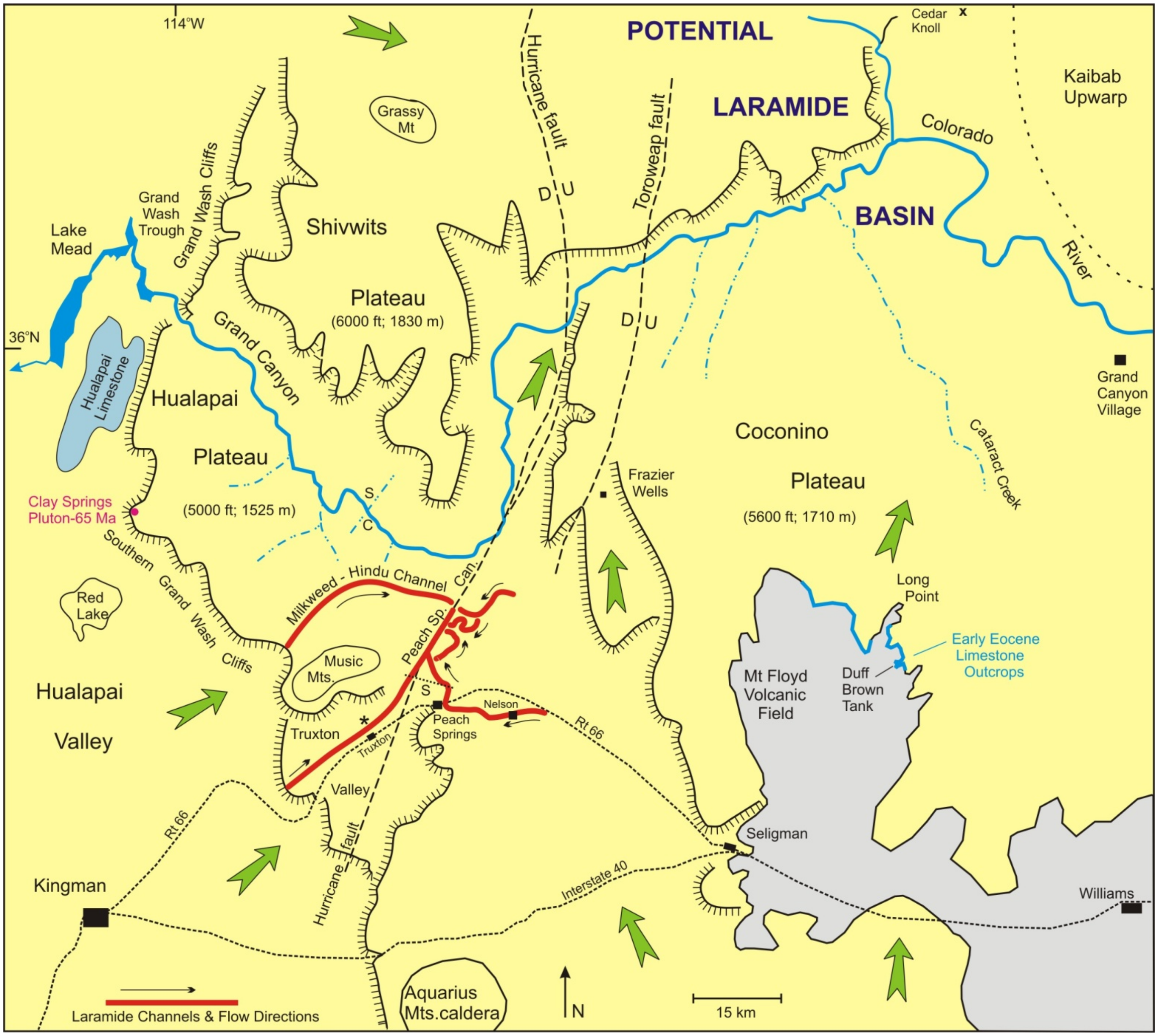

Figure 1. Map of locations mentioned in text. Cross section of figure 2 is at dotted line S. Slate Mountain is located near the word "fault" in "Hurricane fault" at south edge of Truxton Valley. Green arrows indicate measured and inferred directions of flow during Laramide episode of erosion and Rim gravel (Music Mountain Formation) deposition.

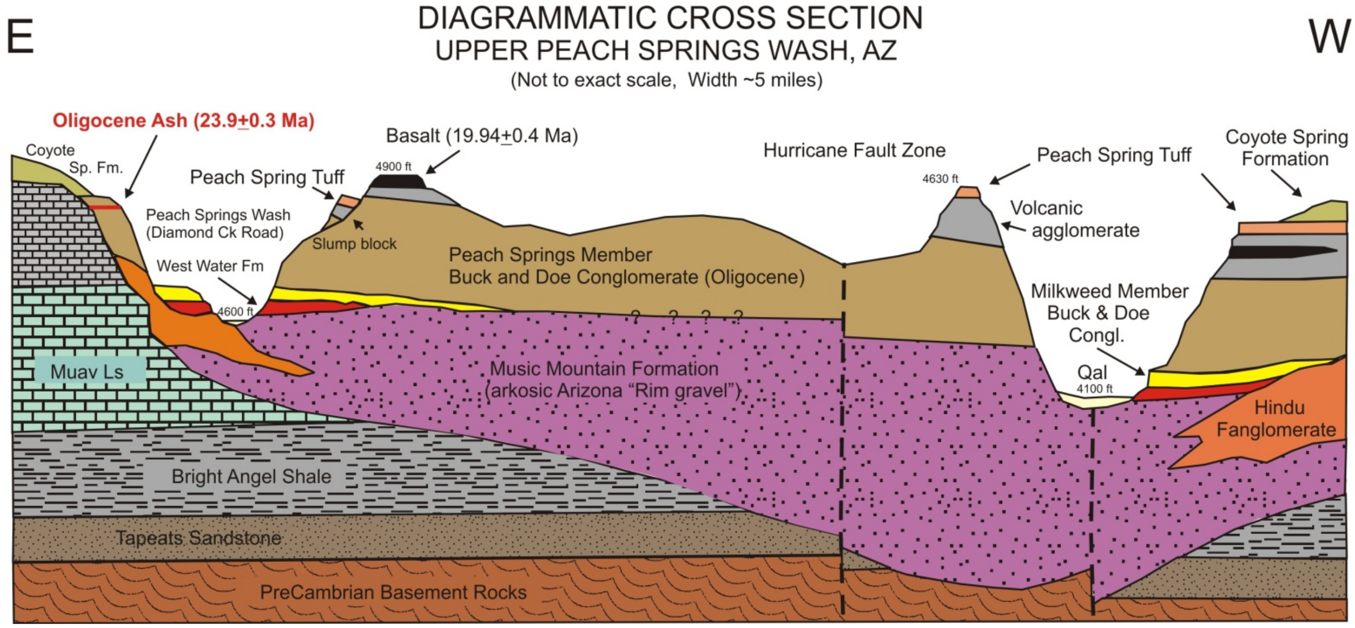

Figure 2. Tertiary stratigraphy of Peach Springs Wash (upper Peach Springs Canyon), Arizona. Note location of newly dated Oligocene ash bed indicated with red font. Formation descriptions and revised Tertiary terminology are described in Young (1999). Presumed Eocene-Oligocene disconformity (depositional hiatus) is at top of West Water Formation (red unit), marked by conspicuous weathered interval shown in figure 4D. Peach Springs Member of Buck and Doe Conglomerate contains high proportion of relatively unweathered exotic igneous clasts from Precambrian outcrops in adjacent Truxton Valley. Geology as originally interpreted by Young (1966). Peach Spring Tuff age is 18.5 ± 0.2 Ma.

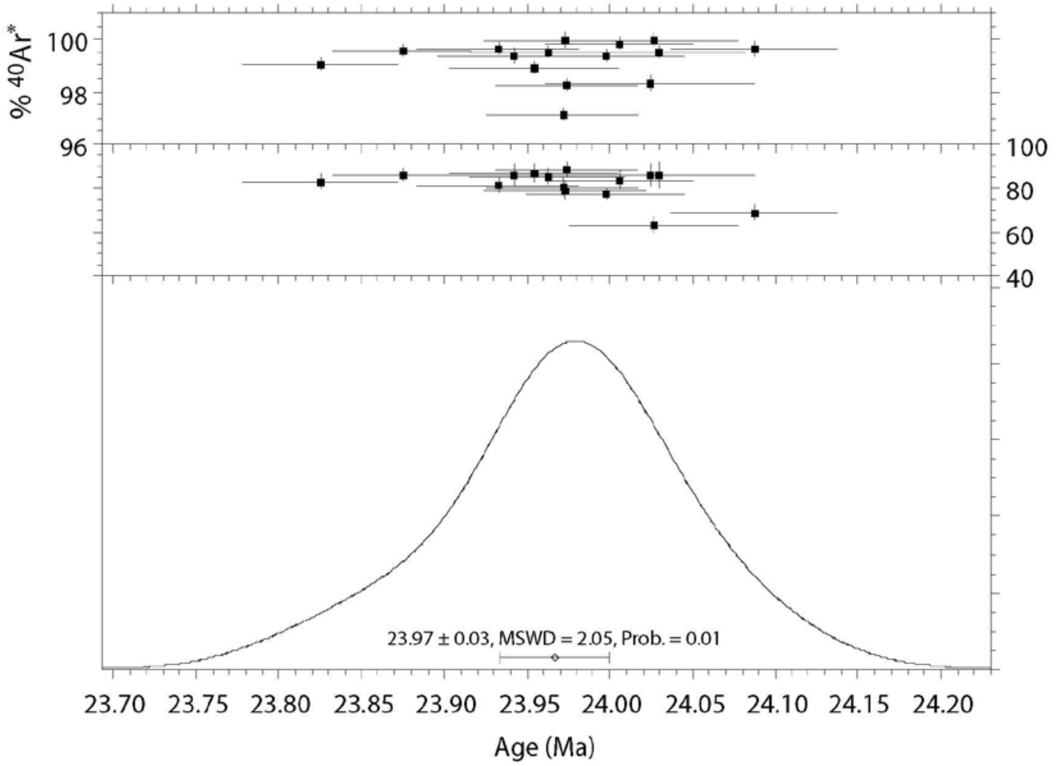

Figure 3. Age-probability plot of single-crystal fusion results from 15 sanidine crystals separated from newly discovered, unnamed Oligocene tuff (1 sigma error bars). ^{40}Ar/^{39}Ar age of 23.97 ± 0.03 Ma is approximately one million years older than the Miocene-Oligocene boundary as revised by Shackleton and others (2000) and now placed at 22.92 ±0.04 Ma. Oldest previously dated basalt on Hualapai Plateau in same Tertiary section has age of 19.94 ± 0.4 Ma (Damon and others, 1996).

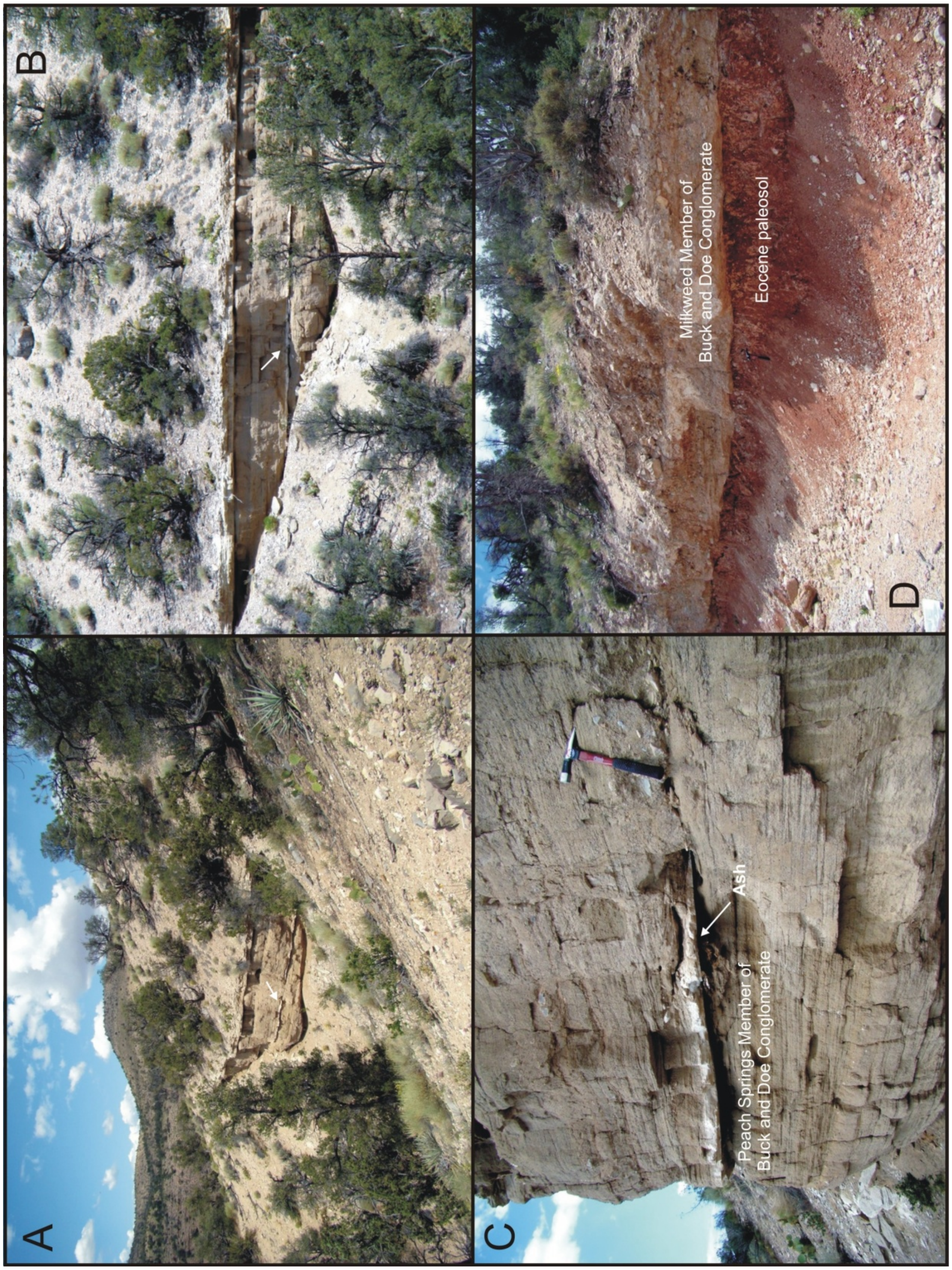

Figure 4. A. Tuff exposure near top of Peach Springs Wash section; view to northeast. B. Tuff outcrop near center indicated by arrow. C. Close-up of Oligocene tuff bed with hammer for scale at same locality as shown in B. D. Disconformable hiatus (nondeposition) between lower Milkweed Member of Buck and Doe Conglomerate and underlying Eocene paleosol (or West Water Formation) approximately 90m below dated ash horizon. Locally derived limestone conglomerate above paleosol records apparent shift to more arid climate.

Early Cenozoic "Rim Gravel" of Arizona—Age, Distribution and Geologic Significance

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Introduction

The northeast-flowing drainage system established in northern Arizona during the Laramide Orogeny was eventually replaced by the southwest-flowing Colorado River following an extended interval of time, which included widespread Basin and Range extension (Young 2001a, 2001b). Our understanding of the events associated with the gradual emergence of an integrated Colorado River drainage system, beginning in latest Miocene time, is hampered by a scarcity of well-dated sedimentary units or volcanic rocks from late Eocene through late Oligocene time, an interval of at least 11 Ma. Most of the volcanic rocks that provide relevant chronologic control adjacent to the plateau margin are Miocene in age or younger; a few are latest Oligocene. Relatively complete Cenozoic sections (Paleocene to Pliocene) are preserved in the partially re-exhumed Laramide paleocanyons (fig. 1) on the Hualapai Plateau (Young, 1989, 1999). These thick canyon fills contain no evidence of either mid-Tertiary incision or regional lake deposits, but do include a prominent disconformity (depositional hiatus), which may represent the climatic deterioration that marks the global Eocene-Oligocene climatic transition (figs. 2, 3). Exotic gravels contained within the basal arkose (Music Mountain Formation, figs. 2,3) should not be confused with the occurrence of much younger widespread exotic lag and reworked gravel deposits located north of Grand Canyon in northern Arizona and southern Utah (Hill and Ranney, 2008). These younger gravels represent one or more generations of reworked clasts from various Paleogene fluvial deposits in Utah, as well as from the Triassic Shinarump Conglomerate (fig. 4C). The reworked gravels north of the Colorado River are associated with the Pliocene and/or Pleistocene incision of the modern Colorado River drainage. These relatively young lag and reworked gravels of northerly derivation must be distinguished from the much older, in situ, immature Tertiary arkoses, which include the so-called Laramide "Rim gravel" of central and northern Arizona.

Rim Gravel terminology

Widespread, poorly exposed, Laramide sediments with limited chronologic control occur in a broad east-west swath stretching across northern and central Arizona from the Hualapai Plateau to the Fort Apache Reservation in eastern Arizona. Many of these Paleogene deposits, originally termed "Rim gravel" for their exposure near the Mogollon Rim, crop out mainly as quartzite-rich, residual lag gravels that blanket their arkosic parent sediments and distort the true clast composition distributions of the intercalated in situ gravel lenses (figs. 4A, B). The Paleogene origin of the Rim gravel was not suspected until the late 1960's and the age was considered by Cooley and Davidson (1963, p. 26) to be as young as Miocene. Cooley and Davidson (1963) are cited by many authors for the definition of "Rim gravel" but Cooley and Davidson attribute the term to McKee (1951), a reference which includes a more comprehensive description of the distribution and stratigraphic position of the exotic gravels, including several Hualapai and Coconino Plateau locations. McKee assumed that the Rim gravels were Pliocene(?), related to presumed Miocene uplift, but he correctly noted that they were located under all the earliest plateau basalts across northern Arizona. Some of these basalts have since been dated as late Oligocene (Young and McKee, 1978). The in situ Rim gravel arkose (fig. 3) mapped on the Hualapai Plateau is now designated as the Music Mountain Formation (Young, 1999), which can be traced discontinuously eastward onto the Coconino Plateau (fig. 2) (Young, 2001; Billingsley and others, 2006). The arkosic sediments were derived from the Laramide highlands that formerly ringed the southern and western margins of the Colorado Plateau. The oldest, weakly lithified, exotic sediments below the Eocene-Oligocene disconformity consist of deeply weathered, fluvial arkosic sands interbedded with subordinate gravel lenses, and isolated thin limestone or marl beds. The most completely preserved Music Mountain Formation sections on the Hualapai Plateau are capped by a thick lateritic soil with well-developed ped structures that clearly marks the disconformity (figs. 3, 4A). This deeply weathered interval may correspond, in part, to the Early Eocene Climatic Optimum (EECO), a global warm episode marked by a paratropical climate from 55 to 53 Ma. Locally derived Oligocene limestone conglomerates directly above the sharply defined disconformity on the plateau remained undated until recently (Young and others, this volume), but they contrast markedly in color, texture, lithology, and provenance with the underlying Laramide arkose.

Revised Age Criteria

Fossil gastropods, charophytes, and ostracodes located in thin limestone beds within arkoses on the Coconino Plateau near Long Point (figs. 1, 2) demonstrate that the age of the associated Paleogene Rim gravel is early Eocene at that location. The fossiliferous limestone beds occur in the upper part of an erosionally truncated arkosic section at Duff Brown Tank, 31 miles (50 km) southwest of Grand Canyon Village (figs. 1, 2). The extent and magnitude of the preserved lateritic soil interval on the Hualapai Plateau suggests that the disconformity might be present elsewhere in the Paleogene landscapes of the Colorado Plateau. However, a corresponding geomorphic record in the region has yet to be widely recognized, with the possible exception of the Eocene surface described in the adjacent southern Rocky Mountains. Global and North American paleoclimate records suggest that the rocks above the Hualapai Plateau disconformity record the transition to early Oligocene aridity. Wider recognition of this disconformity (depositional hiatus) could provide an improved means of unraveling the early and mid Tertiary evolution of the Colorado Plateau landscape. The challenge is to devise a multidisciplinary strategy that can detect the subtle evidence for the Eocene-Oligocene hiatus.

The broad age range of the in situ Laramide gravels (eg., Rim gravel) that still cover Arizona portions of the Colorado Plateau south of the Colorado River and that presumably were shed from marginal uplifts between late Cretaceous and late Eocene time is reasonably constrained by the following independent lines of evidence.

- The well-constrained timing of Laramide tectonism and regional volcanism (Young, 2001a, 2001b). Laramide orogenic activity is the only recognized tectonic event that logically can explain the observed, Late Cretaceous-Paleogene geologic relationships (Damon, 1964; Young, 1979).

- The mean K-Ar age distribution of volcanic clasts (117-51 Ma) contained within Rim gravel sections, which also record progressive erosional unroofing of the source areas (Young, 2001b; Young and Spamer, 2001, Appendix A).
- Closely spaced (<1m) paleomagnetic reversals within a clay-silt fluvial section, which indicate that the Long Point, Arizona, arkosic Rim gravel is younger than the Cretaceous "long normal" Superchron (C34, 120-83 Ma), (Elston and others, 1989; Elston and Young, 1991). These paleomagnetic data record short magnetic reversals and unusual pole positions that best match Chron 24 (circa 54-52 Ma). Chron C24n, which has a relatively short duration of ~1.25 Ma, is subdivided into 5 alternating normal and reversed magnetochron intervals, three of which are as short as 50,000 to 100,000 years in length (Westerhold and Rohl, 2009).
- The complementary chronology preserved in the well dated Paleocene-Eocene sediments of southern Utah (Goldstrand, 1991, 1992, 1994).
- Late Eocene ash beds in the Mogollon Rim Formation (Apache reservation, eastern Arizona) (Potochnik in: Young and Spamer, 2001, Appendix A)
- (U-Th)/He thermochronometry data on the approximate timing of plateau uplift and concurrent Laramide denudation to the Kaibab surface at ca. 50 Ma (Flowers and others, 2008; Kelley and others, 2001).
- The thoroughly weathered in situ condition of true Rim gravel, which contain a wide variety of completely decomposed, exotic, crystalline, Precambrian clasts (Young 1966, 2001a, 2001b).
- The close association of buried paleochannel systems with active Laramide compression (contemporaneous monoclinal deformation and associated ponding) in paleocanyons on the Hualapai Plateau (Young, 1979).
- The location of a 65 Ma pluton (Clay Springs, fig. 1) at the plateau margin unconformably overlain by Miocene basalt flows. The Hualapai Plateau Cambrian strata from the Tapeats Sandstone up through the Muav Limestone were intruded and then extensively eroded to create the existing eroded Hualapai Plateau surface after 65 Ma, but prior to 19 Ma (Young ad McKee, 1978).
- The unconformable relationship between the older, reddish-hued Rim gravel arkose, formed in a paratropical climate, and the younger Oligocene deposits with contrasting arid climate affinities (Hualapai Plateau, Young, 2001a).
- The presence of late Oligocene volcanic rocks covering the regionally stripped surface of Cambrian and Precambrian rocks at the edge of the plateau near the Aquarius Mountains (Young and McKee, 1978).
- The paleontological age of early Eocene lacustrine limestones in the Duff Brown Tank section near Long Point southwest of Grand Canyon as described below (Young and Hartman, 1984; Young 2001b).

Paleontology

The Duff Brown Tank Locality (L4371) limestone beds of the Music Mountain Formation contain a relatively well-preserved assemblage of continental mollusks representing a lacustrine environment of similar age and faunal content to horizons within the lower Eocene Green River Formation (figs. 6a-e). Comparable taxa also are present in the Flagstaff Member/Formation, San Jose Formation, and Wasatch Formation of Wyoming and Utah. Specific biostratigraphic correlations are still under study, but the fauna consists of at least two species of viviparids, two plurocerids, three hydrobioids, a depressed planorbid, a physid, and an ellobiid species (figs. 6a-e). These taxa are based on specimens that are generally undeformed or only mildly distorted and that preserve replaced external shell with good surfaces and sculpture. The discovery of limestone bed occurrences in arkosic sediments near Long Point was originally reported by students of Eugene Shoemaker (Squires and Abrams, 1975), who misidentified some incomplete shells fragments seen in cross section as pelecypods, and assumed a Miocene age for the enclosing sediments (personal communication). Young (1982, 1987) traced the local extent of the limestone beds (fig. 1) during the 1980's over a lateral distance exceeding 11 km. and determined that the thickest limestone section (base covered) exceeds 100 ft (30.5 m), but that most exposures are relatively unfossiferous. The abundant and diverse mollusk assemblage partially illustrated here (fig. 6) is found only at the Duff Brown Tank locality (Young and Hartman, 1984), but many of the thin limestone outcrops elsewhere commonly contain charophytes and pervasive vertical burrows of uncertain affinity. A single ostracod and several stromatolitic algal forms also have been found, as well as some small disarticulated skull fragments, insufficient to identify (E.H. Colbert, personal communication).

The upper Mogollon Rim Formation of eastern Arizona (presumed Rim gravel equivalent) contains ash beds of late Eocene age (37-35 Ma), and a few incorporated volcanic clasts are dated as early Eocene (57-54 Ma) (Potochnik, 2001, Appendix A). This restricts Rim gravel deposition in that part of eastern Arizona to the late Eocene, possibly extending somewhat lower into the middle Eocene. By comparison, fourteen randomly collected volcanic clasts from arkoses collected at Hualapai Plateau and Long Point sections (figs. 1-4) have K-Ar ages of 117 to 64 Ma (Young and Spammer, 2001, Appendix A, p. 244). Five additional (U-Th)/He ages on volcanic clasts from the youngest Long Point section range from 115 to 51 Ma (Flowers and others, 2008). These new (U-Th)/He ages from Long Point are all on clasts collected in the uppermost, volcanic-rich facies (fig. 2) that is presumed to be stratigraphically above the fossil gastropod horizon (Young, 2001b). When combined with the Long Point fossil collections, the available chronology suggests that the relatively unweathered, lighter-hued Mogollon Rim Formation of eastern Arizona is probably a relatively younger facies within the longer ranging Rim gravel sequence, especially when physically contrasted with the more intensely weathered red sediments on the Hualapai and Coconino Plateaus in western Arizona. The much less weathered Mogollon Rim Formation (clasts are relatively durable and competent) is locally underlain by reddish, arkosic sediments, which appear more similar to the weathered arkoses on the Hualapai Plateau. The Rim gravel that is so widespread across parts of central and northern Arizona likely spans a relatively broad interval of time, possibly from latest Cretaceous through late Eocene time. The arkosic gravel appears to become generally younger to the east, as also suggested by the apatite thermochronometry data of Flowers and others (2008).

Plio-Pleistocene Colorado River Basin Gravels

Throughout parts of southern Utah and northern Arizona from eastern Lake Powell westward to St George, Utah, and southward to the Colorado River, there are very different generations of reworked, exotic, quartzite-rich lag gravels on divides and isolated strath terraces associated with the Pliocene incision of south-flowing tributaries of the modern Colorado River (eg., Kanab Creek, Paria River, Escalante River). The locations and exotic nature of many of these reworked gravel deposits are briefly noted in a variety of geologic reports (Phoenix, 1963; Young, 2001;

Billingsley and Priest, 2010) and are compiled by Hill and Ranney (2008). Some of these gravels occur on the Kaibab uplift north and southwest of Jacob Lake at elevations as high as 7600 ft (2316 m). Nearly all of these younger reworked gravels have been shown to contain exotic chert litharenite clasts reworked from the Paleocene Canaan Peak Formation, although this critical lithology, noted by Goldstrand (1992), is not defined or described by Hill and Ranney (2008). Some clasts in these reworked gravels also can be inferred to closely match local exposures of the Shinarump Conglomerate (fig. 4D) (ongoing reconnaissance fieldwork: K. Karlstrom, L. Crossey, T. Hanks, R. Young). The presence of exotic quartzite cobbles, by themselves, in these widespread reworked gravels is insufficient to make precise correlations with the Canaan Peak Formation, because several Paleocene and Eocene lithic units in southern Utah contain gravels dominated by exotic quartzite clasts (Goldstrand, 1992). The original source for the most diagnostic non-quartzitic Canaan Peak clasts is the Mississippian Eleana Formation as described by Goldstrand (1990, 1992, 1994).

Overall, the physical appearances of exotic lag gravel deposits are completely different on the north and south sides of the Colorado River. For example, true Arizona Rim gravel (Music Mountain Formation) that occurs south of the Colorado River in west-central Arizona contains virtually none of the black chert clasts that are otherwise abundant in most exotic reworked gravels north of the Colorado River (fig. 4). In addition, the quartzite clasts derived from southern Utah are more diverse and exhibit a wider range of colors than those exposed in the Music Mountain Formation on the Hualapai and Coconino Plateaus. However, contrary to the assertion in Hill and Ranney (2008), some gravels with Canaan Peak-derived clasts are actually present in strath terrace deposits immediately south of the Colorado River near Marble Canyon. These younger gravels at Marble Canyon are clearly the result of deposition by the emerging Plio-Pleistocene Colorado River when it flowed at elevations near 3600 ft. (1097 m), 500 ft. (152 m) or more above the modern channel. Canaan Peak-derived chert litharenite clasts also occur in modern Colorado River channel deposits at Lee’s Ferry, immediately downstream from Lake Powell. These younger, reworked, lag gravels have an obvious maturity (concentration of very resistant lithologies) that contrasts directly with the marked immaturity of preserved primary (in situ) Rim gravel. The multigenerational, reworked, younger lag deposits north of the Colorado River generally lack any substantial percentage of finer matrix sediment, whereas the true Rim gravel is predominantly weakly lithified arkosic sandstone, with subordinate gravel lenses.

A clear understanding of the origin, field identification, and age of the true (in situ) Arizona arkosic Rim gravel is essential to an appreciation of how the gravel may constrain the evolution of the late Eocene to middle Miocene plateau landscape, and how it may have influenced the eventual evolution of the course of the modern Colorado River. By contrast, the age range of the widely dispersed, multigenerational, compositionally diverse, reworked, resistant lag gravels north of the Colorado River cited by Hill and Ranney (2008) is undoubtedly restricted to the interval of modern Colorado River tributary incision in late Neogene and/or Pleistocene time. Their sources and distribution indicate that these younger reworked gravels track the gradual incision of the modern Colorado River drainage, but shed little light on the frustrating issue of how the Colorado River became integrated from east and west across the Kaibab upwarp.

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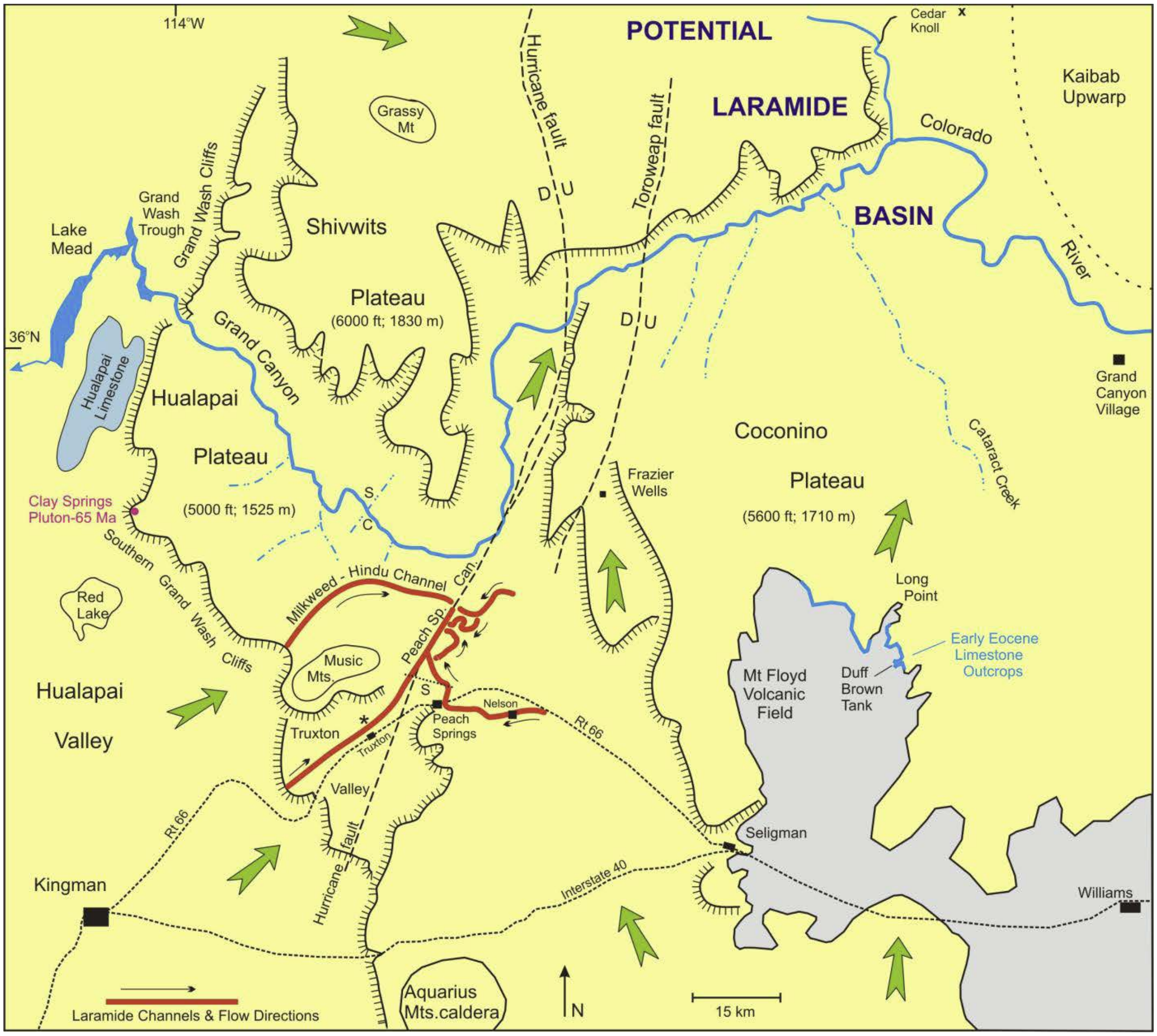

Figure 1. Key localities and places mentioned in text and Laramide paleochannels (red) on Hualapai Plateau. Green arrows indicate measured and inferred directions associated with deposition of Rim gravel (Music Mountain Formation).

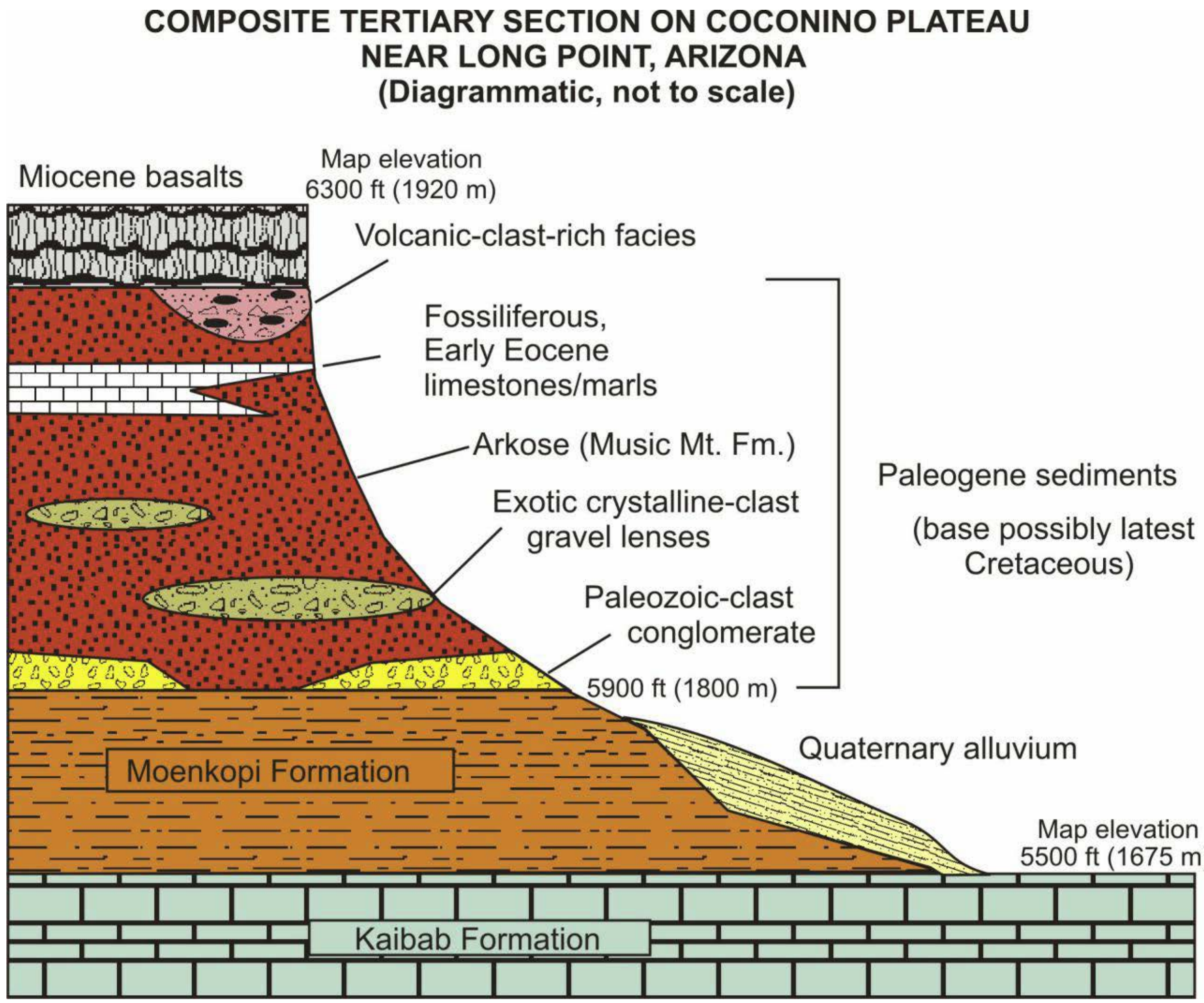

Figure 2. Composite section of Arizona Rim gravel stratigraphy (Music Mountain Formation) near Long Point, Coconino Plateau, Arizona.

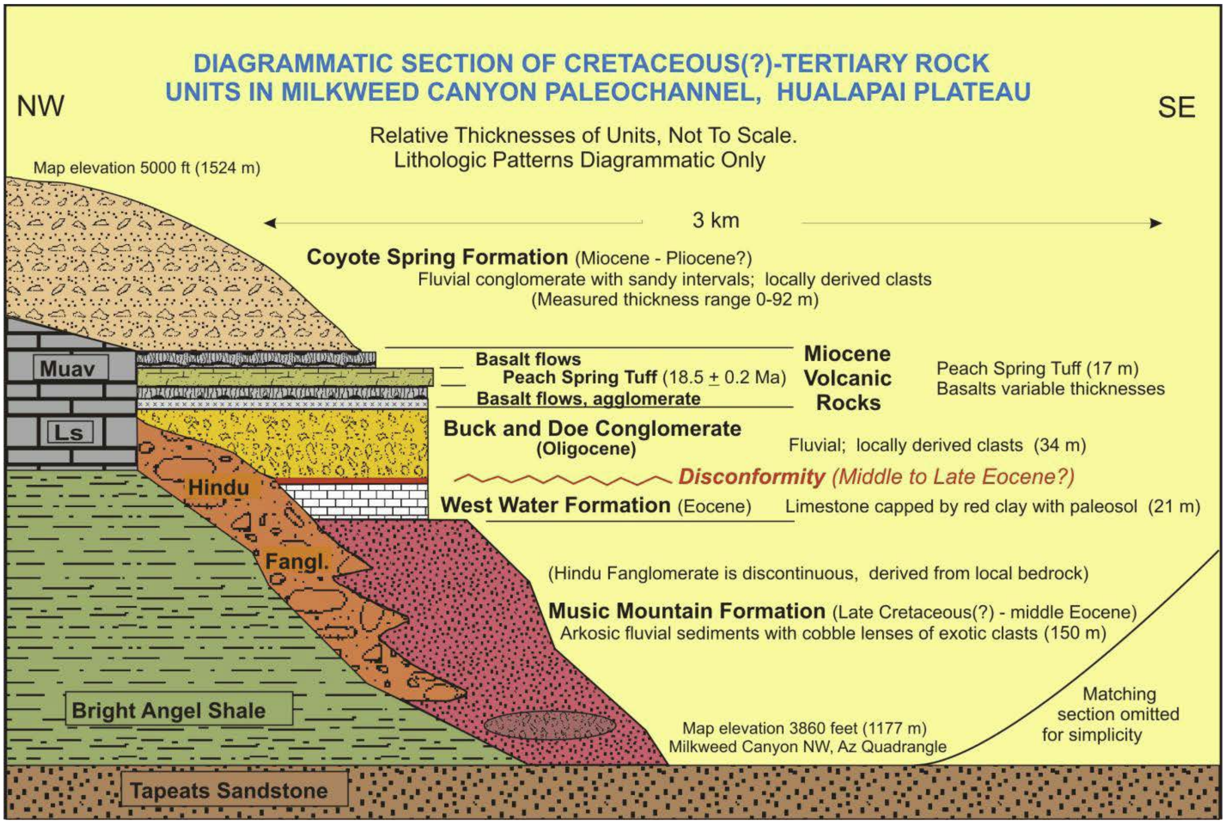

Figure 3. Tertiary stratigraphy of western Hualapai Plateau in Milkweed, Hindu, and Peach Springs exhumed paleocanyons. West Water Formation limestone is presumed to be approximate time equivalent of limestone facies near Long Point, Arizona (fig. 2). New age determination on Buck and Doe Conglomerate in Peach Springs Wash confirms previously estimated Oligocene age for youngest member of Buck and Doe Formation (Young and others, this volume).

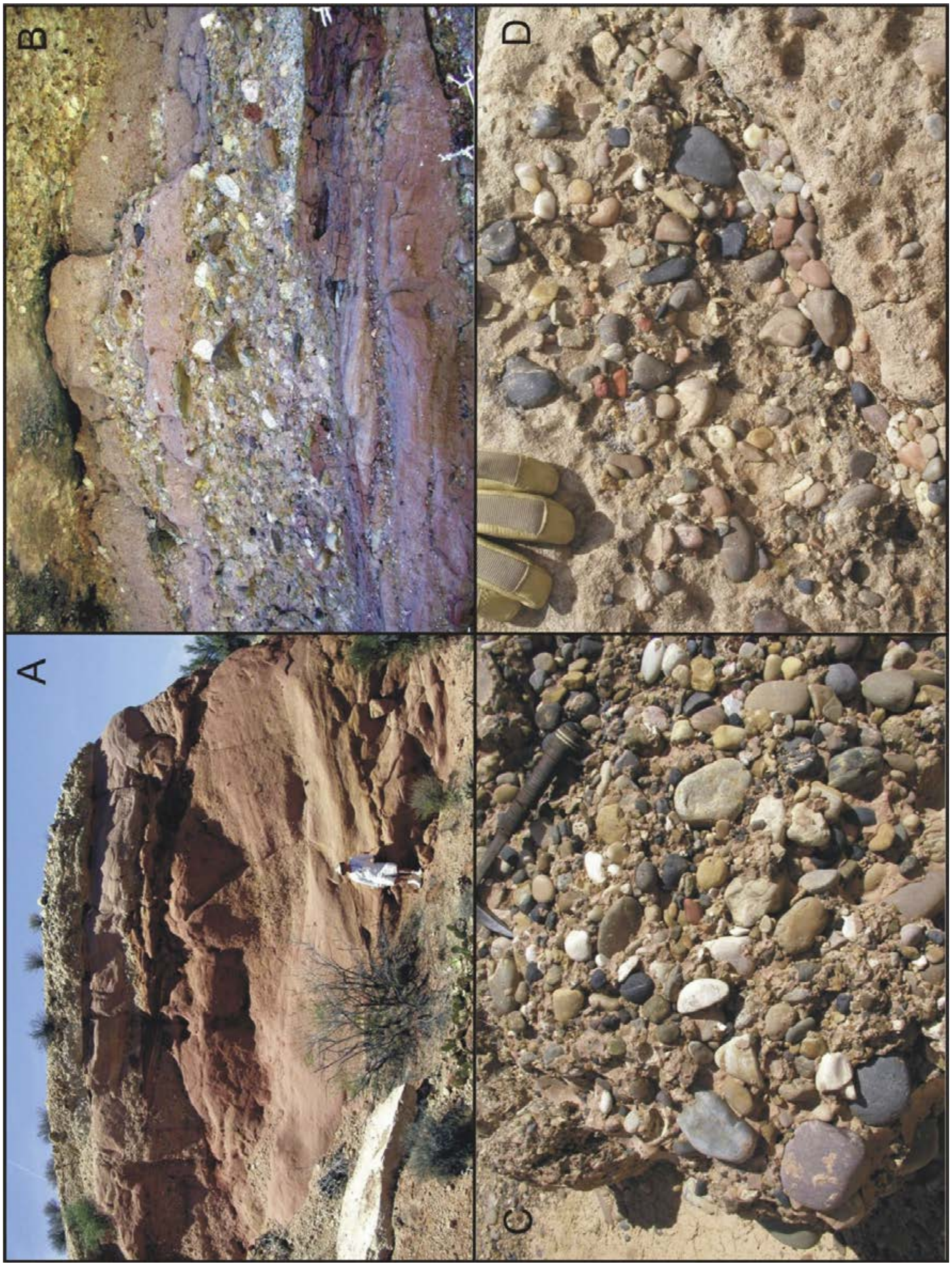

Figure 4. A, B, Rim gravel (Music Mountain Formation) in Peach Springs Canyon, compared with Little Cedar Knoll gravels, C, and in situ Shinarump Conglomerate, D. Note prominence of black cherts and highly resistant clasts in C and D, located north of the Grand Canyon.

Figure 5. Outcrop of early Eocene limestone beds near Duff Brown Tank, Long Point, Arizona. This location produced no mollusks, but abundant charophytes.

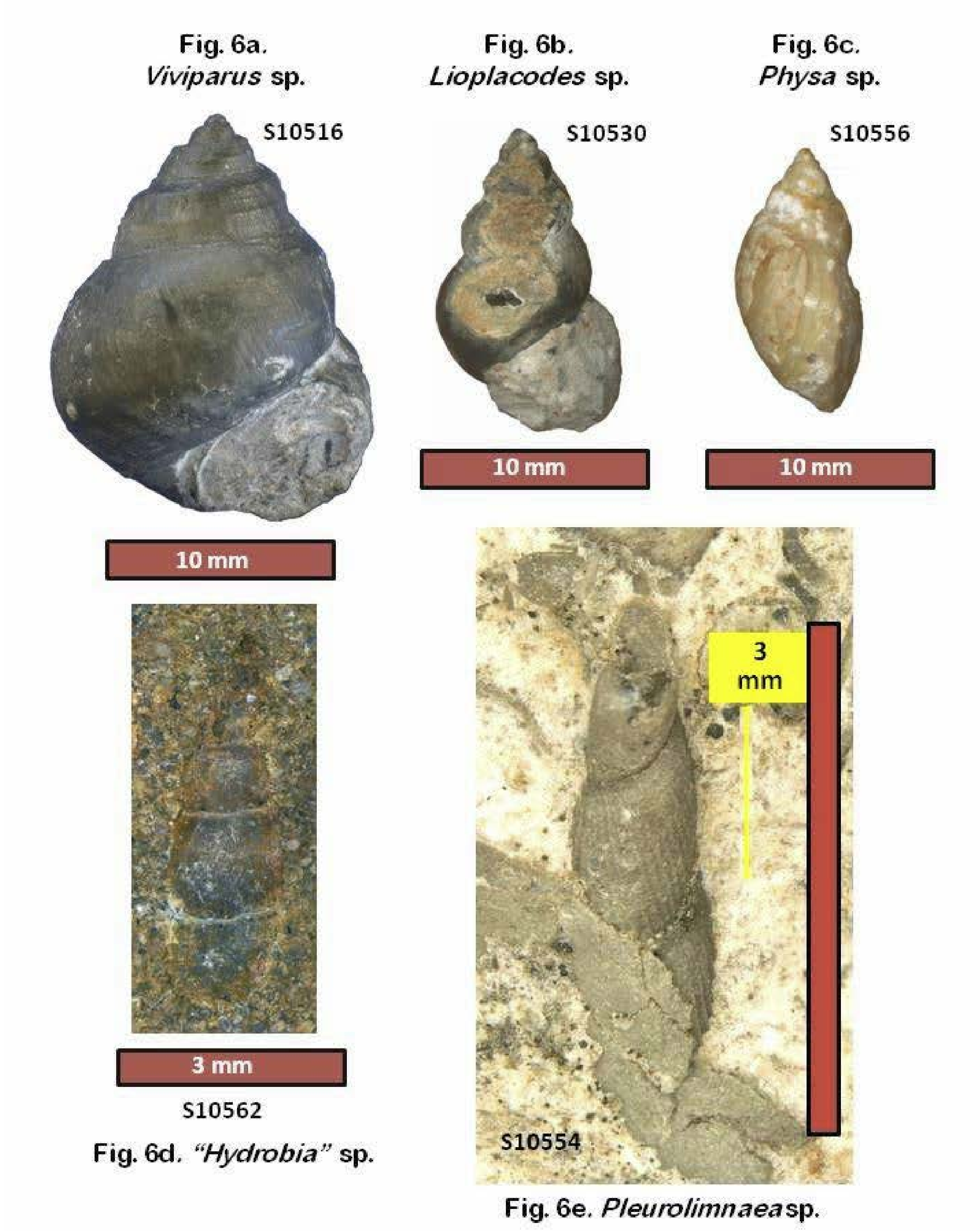

Figure 6. Representative fossil specimens of 5 prominent genera collected at Duff Brown Tank locality (L4371) identified by J.H. Hartman during work in progress on collections of R.A. Young.

Denudational Flexural Isostacy of the Colorado Plateau: Implications for incision rates and tectonic uplift

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INTRODUCTION

The longitudinal profiles of the Colorado River and its tributaries depart from the concave-upward shape of an equilibrium river system and display evidence of geologically recent disturbance. This evidence includes the presence of multiple knickpoints, over-steepened reaches in the Grand Canyon, and high spatial variability of incision rates ranging from ~10-500 m/Ma. These are accompanied by the onset of rapid denudation across the northern half of the plateau since 4-6 Ma and in the southern Rocky Mountains since 8-10 Ma (Kelley et al., 2007; Kelley et al., 2010; Karlstrom et al., 2008; Lee et al., 2010; Hoffman et al., 2010). Two end-member scenarios for this post-10 Ma rapid incision are: 1) that it is being driven by Neogene, and perhaps ongoing tectonic rock uplift, or 2) that uplift occurred during the Laramide, and recent disequilibrium results from a drop in base level associated with drainage integration across the Colorado Plateau. One of the keys to distinguishing between these scenarios may be in the type of rock uplift that is associated with spatial variation in observed incision. If the first scenario involving Neogene uplift is correct, current rock uplift should be tectonic and the magnitude of incision should contain components from both tectonic and isostatic origin, while in the second scenario, where erosion exploits past uplift, current rock uplift and incision should result primarily from the isostatic response to recent exhumation.

In this paper we utilize the basic relationship

$$\Delta E = U_T + U_I - I \tag{1}$$

between rock uplift (U), incision (I), and change in river elevation (ΔE), and divide rock uplift into tectonic (U_T) and isostatic (U_I) components. The isostatic component refers only to the isostatic response to crustal unloading by erosion. All other mechanisms of rock uplift are lumped into the tectonic component. According to Eqn. (1), incision (I) minus isostatic uplift (U_I) gives a "residual incision" not accounted for by isostacy that is equivalent to the difference between tectonic uplift and change in river elevation.

The goals of this paper are to: 1) estimate the net thickness of bedrock eroded from the Colorado Plateau and nearby southern Rocky Mountains since 10 Ma; 2) compute the flexural isostatic response to this denudation; 3) remove the isostatic response from observed magnitudes of incision to produce a "residual incision"; and 4) interpret the residual incision in terms of tectonic uplift using a scenario for change in river elevation. A 10 Ma paleo-surface was chosen as an isochronous reference for measuring incision in this study because it: 1) predates the onset of rapid incision in the southern Rocky Mountains and Colorado Plateau (Kelley et al., 2007; Aslan et al., 2010); 2) predates integration of the Colorado River to the Grand Wash Cliffs ~6 Ma (Karlstrom et al., 2008); 3) postdates most development of the Basin and Range; 4) postdates erosion of the Chuska Erg on the southern portion of the plateau (Cather et al.,2009); and 5) postdates denudation of the Grand Canyon region to the Kaibab Limestone surface (Flowers et al.,2009).

This paper builds on previous attempts to quantify isostatic response to erosion of the Colorado Plateau (Pederson et al., 2002a, 2002b, 2007; Pederson, 2006; McMillan et al., 2006; and Roy et al., 2009). These studies involved different time periods and different areal extent than the present study. Pederson et al. (2002a) used the present elevation of coastal marine sediments deposited during regression of the Cretaceous Interior Seaway to determine net tectonic uplift and the isostatic response to net erosion since 65 Ma. They found that removing isostatic rebound from total rock uplift leaves an average of 1800 m of residual uplift across the entire Colorado Plateau to be explained by mechanisms other than erosional isostacy. Pederson (2006), and Pederson et al. (2007) extended their work to include flexural isostacy and discovered a "bulls eye" of more than 1 km of isostatic rebound in Canyonlands, and postulated a contribution of this rebound to high short term incision rates inferred from cosmogenic surface dates in that region (e.g. Cook et al., 2009). Overall, Pederson et al. (2002a) favored a scenario in which early Cenozoic tectonic events (Laramide Orogeny) provided most of the uplift of the plateau, with only passive isostatic response to denudation since that time (scenario 2 above).

The approach taken by McMillan et al. (2006) was to define recent erosion of the Colorado Plateau, southern Rocky Mountains, and western Great Plains by reconstructing the aggradational surface of mid-to-late Cenozoic basin fill deposits, and to measure subsequent incision relative to this surface. The transition from deposition to erosion of these formations generally occurred between 5-10 Ma so their "datum" was not isochronous and was much younger than that used by Pederson et al. (2002a). The overall conclusions of McMillan et al. (2006) were that the Colorado Plateau experienced slow subsidence during Oligocene-Miocene time of ~850 m that accommodated widespread aggradation, and that rapid incision has been initiated by tectonic uplift since ~8 Ma (scenario 1 above).

Roy et al. (2009) used the Pederson et al. (2002a) rock uplift model from which they computed and removed the isostatic component of uplift due to both net Cenozoic erosion and extensional unloading to obtain a residual rock uplift of 1.6 km in the central plateau and 1.9 km near the margins. They modeled this residual as a response to conductive heating of the lithosphere by volcanic intrusions and progressive thinning of the lithosphere of the Colorado Plateau from the perimeter toward the center. Their uplift was initiated ~30 Ma and would be ongoing since thermal equilibrium has not been achieved. While this model involves recent and ongoing uplift (scenario 1), the extended uplift period (30 Ma), slow uplift rate, and small magnitudes of uplift are not consistent with the onset of rapid incision since 10 Ma in the southern Rocky Mountains or since 4-6 Ma in the Canyonlands region. Furthermore, conductive geodynamic models seem less applicable than convective models (Moucha et al., 2009; van Wijk et al., 2010), especially given evidence for convective flow documented by patterns and geochemistries of basaltic magmatism (Crow et al., 2010).

METHODS

The studies by Pederson and Roy treated the Colorado Plateau as an isolated entity, yet there is no known tectonic decoupling between the Colorado Plateau, southern Rocky Mountains and western Great Plains, and they are unified in their response to epeirogenic thermal and dynamic uplift associated with shallow asthenosphere below the southern Rocky Mountains and Grand Canyon region (Eaton, 2008; Dueker et al. 2001; Aster et al., 2009). In this paper we follow McMillan et al. (2006) in using a broad region extending from the eastern plains of Colorado to eastern Nevada, and from southern Wyoming to southern Arizona. Using a broad region for analysis removes edge effects of the ~160 km radius isostatic rebound filter from the region of interest on the Colorado Plateau and southern Rocky Mountains.

Reconstruction of the 10 Ma surface

Constraints used to reconstruct present elevations of the 10 Ma surface consist of 8-12 Ma basalts that preserve remnants of the surface, and AFT(He/Th) thermochronometry data that provide estimates of timing and thickness of eroded sediments in areas where basalt constraints are sparse, and where all traces of the original surface have been removed by erosion. Figure 1 shows the location of the study area and an image of topographic elevation with the distribution of control points used to constrain the 10 Ma paleo-surface. Black lines indicate state boundaries, white lines indicate boundaries of the Colorado Plateau and southern Rocky Mountains, and circled numbers define regions within the control points that are discussed in detail in the full paper, but will not be presented in this extended abstract in the interest of brevity. Some of these areas are briefly described in the caption for Figure 1. Black control points indicate basalt, red points are thermochronometry locations, and white points are additional constraints that define topographic transitions or fill regions of known elevation to allow for smooth interpolation of the surface using triangular interpolation methods. All together there are 377 control points.

The primary source for basalt constraints is the NAVDAT database (http://navdat.org) where dated basalt flows were extracted within geographic coordinates -116W to -102W and 31N to 42N using an age range from 8-12 Ma and resulted in 90 unique data points. Additional basalt data for the southern Rocky Mountains was from Aslan et al., (2010) and Cole et al., (2010). Numerous 10 Ma basalt flows are found on the perimeter of the Colorado Plateau and in the Basin and Range, and are used to define the transition in elevation from the plateau to the Basin and Range in the paleo-surface. The southern plateau (Chuska erg) had been eroded to near present elevations by 10 Ma (Cather et al., 2009), and thermochronometry data indicate that the Grand Canyon region had been denuded to the present Kaibab limestone surface with an escarpment of ~1-2 km thick Mesozoic section extended from just north of the Grand Canyon region to south of Lee's Ferry, then to the Chuska Mountains and over to the northern extension of the Rio Grande rift on the eastern margin of the plateau. The Mesozoic section was buried by thick Cenozoic deposits across the NW plateau and extending eastward into the Southern Rocky Mountains (Kelley et al., 2007; Kelley et al., 2010; Lee et al., 2010; Hoffman et al., 2010). These data are particularly useful in the Canyonlands region where there are no remnants of the 10 Ma surface, but they have large uncertainties (factor of 2) that can be used to bracket the range of feasible isostatic solutions.

Solving for net eroded thickness since 10 Ma

Triangular interpolation of the control points shown in Figure 1 was used to create the predicted present day elevation of the 10 Ma paleo-topographic surface after isostatic rebound. Subtraction of present topography from the predicted surface gives a net eroded thickness since 10 Ma of 1250 - 1500 m along the Colorado River corridor from Grand Junction, Colorado through the Grand Canyon (Figure 2). In the upper reaches of the Colorado River, and in the Grand Canyon, incision is confined to narrow canyons, while in the region of Canyonlands National Park denudation has been widespread and is strongly correlated with highly erodible lithologies that include the Morrison Formation and Mancos shale. The post- 10 Ma erosion is concentrated in the northern half of the Colorado Plateau and the edge of the Great Plains since erosion of the Chuska Erg on the southern plateau occurred between 30-16 Ma (Cather et al., 2009), and denudation of the Grand Canyon region to the Kaibab Ls. was also prior to 10 Ma (Flowers et al., 2008). Figure 2 is analogous to Figure 7 in Pederson et al. (2002a) and has similar eroded thickness for the northern portion of the plateau, even though this denudation took place in the last 10 rather than 30 million years. This is compatible with other geologic evidence that denudation rates were very low between 30-10 Ma in this region (Larson et al., 1995 ; Aslan et al., 2010).

Solving for Flexural Isostatic Rebound

The flexural isostatic response to erosional unloading of the crust was computed using a Fourier transform based algorithm. The procedure is to compute a 2-dimensional fast Fourier transform (FFT) of the eroded thickness given in Figure 2, multiply the wavenumber spectrum by the Fourier transform of the flexural isostatic operator that converts crustal load into deflection of the lithosphere, and perform the inverse 2-dimensional FFT to obtain the isostatic rebound displayed as contours in Figure 2. We chose effective elastic thickness for the lithosphere of 25 km based upon the coherence analysis of Lowery et al. (2000). We also used Young's Modulus of 70 GPa, Poisson's Ratio of 0.25, density of the crust of 2650 kg/m^3, and mantle density of 3300 kg/m^3 based upon values from Turcotte and Schubert (2002), giving a flexural rigidity of 10^{23} N-m. The white contours in Figure 2 represent flexural isostatic rebound with a 100 m contour interval. Maximum rebound of 1100 m is centered over the confluence of the Green and Colorado Rivers, and the response is dome shaped and diminishes to ~200 m near the perimeter of the Colorado Plateau. This is very similar to the "bulls eye" found by Pederson et al. (2007) using a 30 Ma time span.

IMPLICATIONS FOR INCISION AND TECTONIC UPLIFT

Figure 2 shows net eroded thickness since 10 Ma and the associated flexural isostatic response, and a cross-section of this response taken along the Colorado River corridor is shown in Figure 3. The black curve is the outer envelope of total incision (corresponding to river level), the light gray represents total flexural isostatic rebound from denudation, and the red curve is residual incision (incision - rebound) not accounted for by isostatic rebound. According to equation 1 this residual is equivalent to the tectonic uplift since 10 Ma minus any change in river elevation during that period. Without paleo-elevation data that would pin down the change in river elevation, it is not possible to determine an accurate magnitude for tectonic uplift. However, we can infer differential tectonic uplift assuming that broad surface uplifts cause rivers to change elevation smoothly along their profiles. We interpret the residual incision as indicating that isostatic rebound accounts for a larger portion of the total incision in the central Colorado Plateau than in the Grand Canyon region or in the southern Rocky Mountains, and we attribute the ~500 m of excess residual incision in those areas to differential tectonic uplift relative to the central plateau.

Below the graph of residual incision is a tomographic image of the upper mantle (Schmandt et al., 2010) showing low velocity Vp anomalies (red), corresponding to hot and buoyant upper mantle, underlying the Grand Canyon region and southern Rocky Mountains and correlating with areas of highest residual incision. Likewise, high velocity Vp anomalies (blue), corresponding to colder and less buoyant lithosphere, underlie the central Colorado Plateau and correlate with the region of lowest residual incision. We interpret this as evidence for a link between upper mantle convective thermal anomalies and surface tectonic uplift in mid-continent orogenic settings.

CONCLUSIONS

The northern Colorado Plateau has been denuded over the past 10 Ma by more than 1500 m of incision that would theoretically produced a dome shaped isostatic rock uplift exceeding 1 km in the vicinity of the confluence of the Green and Colorado Rivers and a surface lowering of ~300 m, both declining toward the perimeter of the plateau. The denudation history of the Southern Rocky Mountains is considered here to be linked to that of the CP and is somewhat constrained based on thermochronologic data. These data broaden and move the "bulls eye" of rebound eastward relative to published models.

Removal of the isostatic component of rock uplift from observed magnitudes of incision can provide estimates of relative tectonic rock uplift along the river profile. Whereas isostatic rebound can perhaps explain post-10 Ma incision magnitudes in the central Colorado Plateau, post-10 Ma tectonic uplift components seem to be required in the Colorado Rockies and western Colorado Plateau. These areas appear to overlie low velocity, buoyant upper mantle, suggesting the surface uplift may be driven by mantle upwelling in these regions. This evidence supports recent and ongoing uplift as the driver of rapid incision of the Colorado Plateau and adjacent southern Rocky Mountains since 10 Ma (scenario 1).

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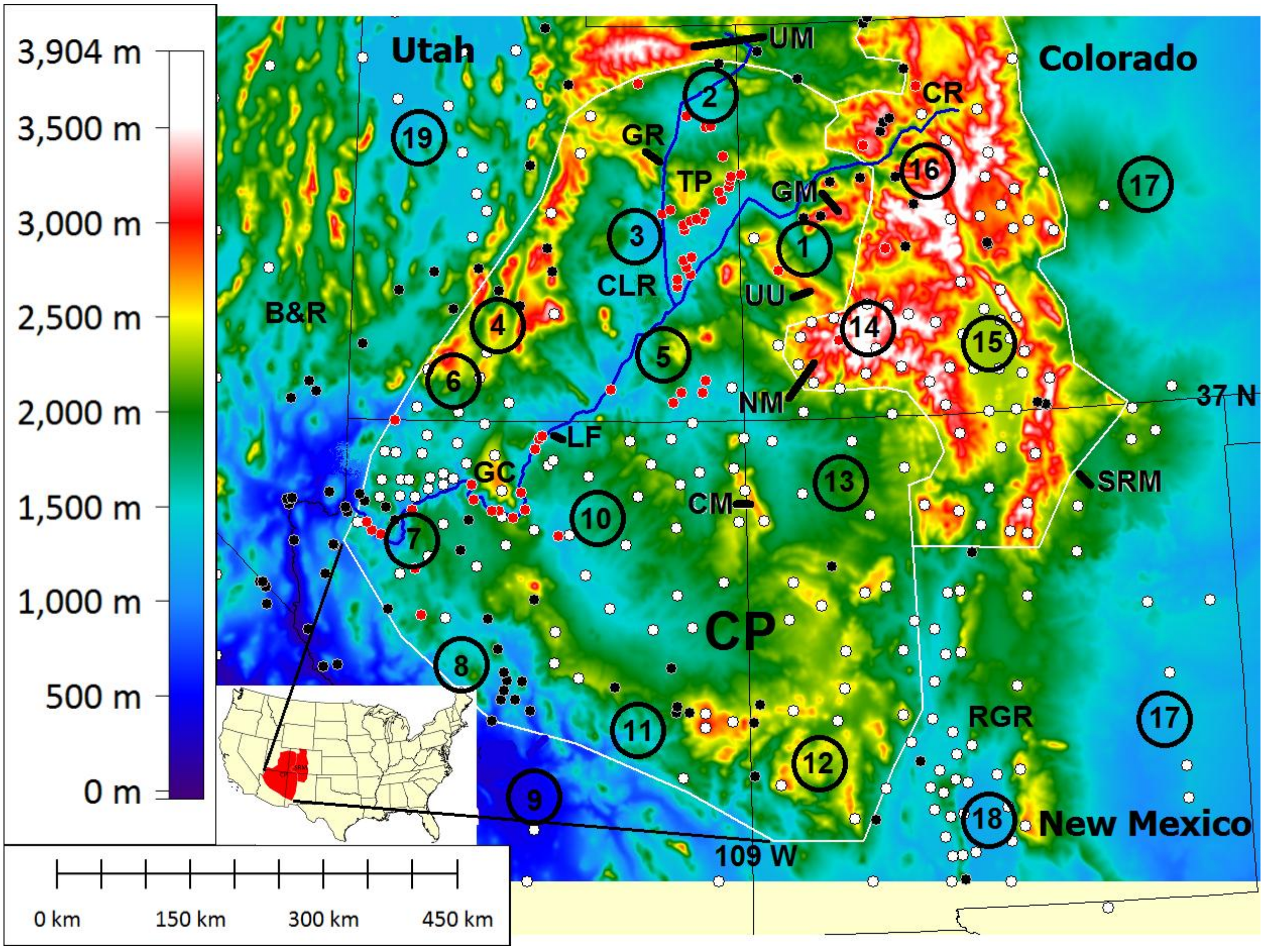

Figure 1 - Topographic image of the Colorado Plateau (CP), Southern Rocky Mountains (SRM), and Basin and Range (B&R) showing control points used to define the 10 Ma paleo-surface. The blue line is the Green and Colorado Rivers, black dots are 8-12 Ma basalt flows, red dots are thermochronometry data points, and white dots are additional elevation constraints that define topographic transitions or fill in regions of known elevation to aid the triangular interpolation algorithm in producing a smooth surface. They are used to help define the volcanic edifice in the Needle Mountains (NM), the eroded basin in which Hopi Lake formed (10), the San Luis Valley (15) and Rio Grande rift (18), the retreating Mesozoic escarpment north of Grand Canyon (6), and high erosional remnants on the eastern plains (17). Other points of interest in the figure are CLR = Canyonlands region, CM = Chuska Mountains, CR = Colorado River, GC = Grand Canyon region and East Kaibab monocline, GM = Grand Mesa, GR = Green River, LF = Lee's Ferry, TP = Tavaputs Plateau, UM = Uinta Mountains, UU = Uncompahgre Uplift.

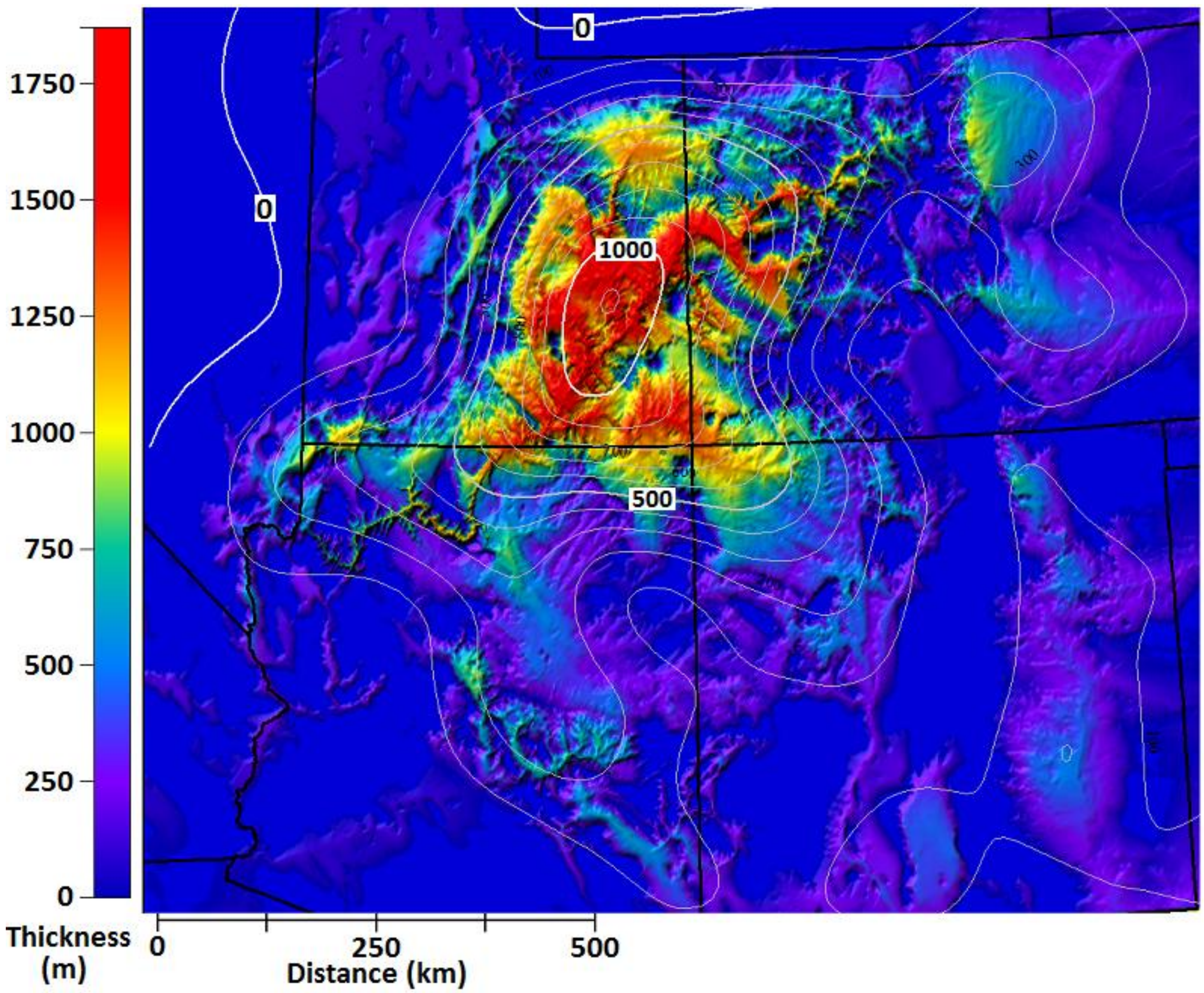

Figure 2 - Estimated eroded thickness for the past 10 Ma and contours of isostatic rock uplift (100 m interval) resulting from erosional unloading of the Colorado Plateau over the past 10 Ma. The Grand Canyon produces little isostatic response (~200 m) due to its narrow extent, while similar eroded thickness in Canyonlands National Park has produced over 1 km of rebound. The region of widespread denudation correlates with highly erodible lithology of the Morrison Formation and Mancos shale, while narrow canyons are incised into more competent Paleozoic and Precambrian rock.

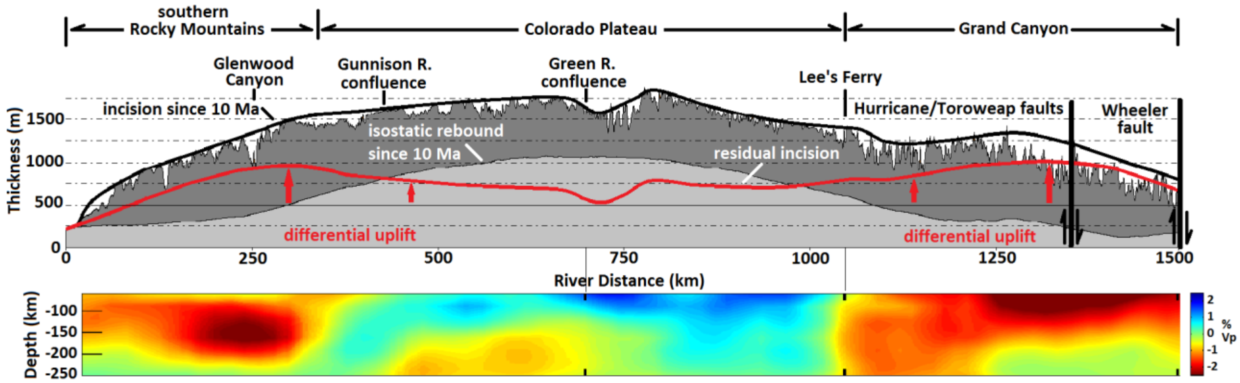

Figure 3 - Residual incision and its correlation with tomographic imaging of the upper mantle. The black curve is the upper envelope of estimated total incision (dark gray) which equals eroded thickness since 10 Ma along the Colorado River corridor; light gray is the associated isostatic rebound; and the red curve is the residual incision = incision - rebound. Isostatic rebound accounts for a larger portion of observed incision in the center of the Colorado Plateau than in the Grand Canyon or southern Rocky Mountains and leaves up to 500 m of differential residual incision relative to the central plateau. This suggests that the residual is due to a tectonic uplift component in those areas. Qualitatively the highest residual incision correlates with low compressional velocity (Vp) in the mantle, and the lowest residual incision correlates with high Vp in the mantle.

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Appendix B. Agenda

Workshop Agenda—May 24-26, 2010, Flagstaff, Arizona
Origin and Evolution of the Colorado River System II

Symposium organizers
Dick Young, State University of New York, Geneseo
Karl Karlstrom, University of New Mexico
Joel Pederson, Utah State University, Logan
Kyle House, Nevada Bureau of Mines and Geology, Univ. of Nevada, Reno
Andres Aslan, Mesa State College, CO
Sue Beard, USGS, Flagstaff
George Billingsley, USGS, Flagstaff

SCHEDULE

Day 1 (Morning): Monday, May 24 - Plenary discussion: workshop goals, research problem

7:30-8:15: Sign-in, Building 3 East lobby

The goal of the first half day session is to formulate and clarify the main research questions and problems for the three main reaches of the Colorado River. These speakers will NOT be allowed to give GSA-style talks.

All talks are 5 minutes followed by an additional 5 minutes for discussion

8:15-8:30 Welcome, logistics and creation of a CR-CP database

Workshop goals, research problems – George Billingsley and Ivo Lucchitta convening

8:30-8:40 Dick Young: Canyons old and Young - Issues & progress since 2000 workshop
8:40-8:50 Brian Wernicke: California paleoriver
8:50-9:00 Karl Karlstrom: Regional mantle-driven uplift
9:00-9:10 Robert Xavier: Mantle-driven uplift
9:10-9:20 Kyle House: Lower Colorado River integration controversies
9:20-9:30 Becky Dorsey: Opening of the Gulf of CA and the first CR sediments to the sea

9:30-10:30 Break and posters

10:30-10:40 Joel Pederson: The Colorado Plateau bullseye
10:40-10:50 Sharie Kelley: Overview of thermochronology of the Colorado Plateau-Rocky Mountain region
10:50-11:00 Kelin Whipple: The Lees Ferry knickpoint
11:00-11:10 Will Ouimet: Slope-Area analysis of channel steepness
11:10-11:20 Andres Aslan: Upper Colorado River - summary of problems and goals

11:20-12:00 General discussion – organizing committee convening

12:00-1:30 Lunch and posters at USGS*

Day 1 (Afternoon): Monday, May 24 – Lower Colorado River and tributaries

Pre-6 Ma drainages and landscapes and opening of the Gulf of CA - Becky Dorsey convening

1:30-1:40 Davis: Another California paleoriver
1:40-1:50 Kris McDougall: Fossil evidence from the Gulf
1:50-2:00 Dave Kimbrough: Detrital zircon constraints

Lower Colorado River integration, lakes, and incision history – Jon Spencer and Keith Howard convening

2:00-2:10 Jon Spencer: Lake spill over
2:10-2:20 Laura Crossey: Lakes and carbonates
2:20-2:30 Paul Umhoefer: Lake Mead area
2:30-2:40 Keith Howard: Lower Colorado River deposits
2:40-2:50 Tracey Felger/Lee Amoroso: Pliocene sediments and basalts
2:50-3:00 Phil Resor: Monoclines

3:00-4:00 Break and posters

4:00-5:00 Discussion and additional input - Kyle House convening

5:00-8:00 BBQ, posters and informal discussions at USGS**

Day 2 (Morning): Tuesday, May 25 - Grand Canyon, Little Colorado River

Laramide through Miocene (pre 6 Ma) drainage evolution and tectonic chronology of the Colorado Plateau – Steve Reynolds and Dick Young convening

8:00-8:10 Joe Hartman: Fossil evidence for age of rim gravels
8:10-8:20 Sue Beard: Kingman arch, paleovalleys
8:20-8:30 John P. Lee: Thermochronology
8:30-8:40 Ron Blakey: Verde Valley
8:40-8:50 Ivo Lucchitta: Crooked ridge
8:50-9:00 Carol Hill: Paleogeography
9:00-9:10 Dick Young: Oligocene ash

Basin integration and incision history - Joel Pederson and Ivo Lucchitta convening

9:10-9:20 Ivo Lucchitta: Muddy Creek, 'the immovable object'
9:20-9:30 Joel Pederson: Groundwater integration
9:30-9:40 George Billingsley: Grand Canyon points of interest
9:40-9:50 Marty Grove: DZs above and below the Lees ferry knickpoint
9:50-10:00 Jessica Lopez Pearce: Evidence from the Hualapai Limestone

10:00-10:30 Break and posters

10:30-10:40 Christopher Tressler: Rock strength and knickpoints
10:40-10:50 Tom Hanks: Incision rate, Glen Canyon
10:50-11:00 Ryan Crow: Incision rate summary
11:00-12:00 Discussion and additional input - Karl Karlstrom convening

12:00-1:30 Lunch and posters at USGS*

Day 2 (Afternoon): Tuesday, May 25, 2010 – Upper Basin Colorado, Green, San Juan, LCR

Pre-10 Ma topographic evolution, thermochronology and uplift of the Rockies – Shari Kelley and Steve Cather convening

1:30-1:40 Markella Hoffman: Thermochronology
1:40-1:50 Charles Ferguson: Powder Rim Gravel
1:50-2:00 Steve Cather: Chuska erg
2:00-2:10 Andre Potochnik: Upper Salt River
2:10-2:20 Bill Dickinson: Bidahochi Formation
2:20-2:30 John Douglass: Integration mechanisms
2:30-2:40 Embid: Little Colorado incision rates
2:40-3:00 Open for additional speakers

3:00-3:30 Break and posters

Basin integration and incision history – Andres Aslan and Karl Karlstrom convening

3:30-3:40 Greg Lazear: Post-10 Ma isostatic response
3:40-3:50 Dave Marchetti: Fremont river
3:50-4:00 Rex Cole: Grand Mesa
4:00-4:10 Andy Darling: Incision rates on Green and Colorado
4:10-4:20 Magdalena Sandoval: Black Canyon and post 640 Ma
4:20-4:30 Open for additional speakers

4:30-5:30 Discussion and additional input - Andres Aslan convening

6:00-8:00 Pizza, posters and informal discussions at USGS***

Day 3 (Morning): Wednesday, May 26, 2010 – Synthesis

8:00-10:00 Discussion and synthesis – Dick Young and Karl Karlstrom convening

10:00-10:30 Break

10:30-12:00 Organize writing teams - further discussion

12:00-1:30 Lunch (ordered in house)

Day 3 (Afternoon): May 26, 2010 – Synthesis and Writing Team

1:30-5:00 Writing team creates outline version of interim report (volunteers encouraged).

The goal is to create an Introductory paper for an electronic Geosphere theme issue entitled: "Origin and Evolution of the Colorado River System II". Papers from workshop participants (and others from the community) will be solicited for the volume for 12 months after the workshop. Multiple co-authors (including conveners and participants) are welcome to participate in the introductory paper and the database that results from the meeting. All work will be carefully attributed to the scientists that contribute the data.

1:30-5:00 Optional - voluntary work on Database of Geochronology, Thermochronology and Incision Data for the Colorado River System (more details coming on Google Earth-linked data base from Kyle House)

***** Lunch, Monday and Tuesday: catered; includes sandwiches, side salad, drinks and cookies; $13 per person per day.

****** BBQ, Monday night: prepared by USGS, will include food and beverages; $12 per person.

******* Pizza, Tuesday night: $12 per person.