The Geological Society of America
Special Paper 439
2008

Stratigraphic evidence for the role of lake spillover in the inception of the lower Colorado River in southern Nevada and western Arizona

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ABSTRACT

Late Miocene and early Pliocene sediments exposed along the lower Colorado River near Laughlin, Nevada, contain evidence that establishment of this reach of the river after 5.6 Ma involved flooding from lake spillover through a bedrock divide between Cottonwood Valley to the north and Mohave Valley to the south. Lacustrine marls interfingered with and conformably overlying a sequence of post–5.6 Ma fine-grained valley-fill deposits record an early phase of intermittent lacustrine inundation restricted to Cottonwood Valley. Limestone, mud, sand, and minor gravel of the Bouse Formation were subsequently deposited above an unconformity. At the north end of Mohave Valley, a coarse-grained, lithologically distinct fluvial conglomerate separates subaerial, locally derived fan deposits from subaqueous deposits of the Bouse Formation. We interpret this key unit as evidence for overtopping and catastrophic breaching of the paleodivide immediately before deep lacustrine inundation of both valleys. Exposures in both valleys reveal a substantial erosional unconformity that records drainage of the lake and predates the arrival of sediment of the through-going Colorado River. Subsequent river aggradation culminated in the Pliocene between 4.1 and 3.3 Ma. The stratigraphic associations and timing of this drainage transition are consistent with geochemical evidence linking lacustrine conditions to the early Colorado River, the timings of drainage integration and canyon incision on the Colorado Plateau, the arrival of Colorado River sand at its terminus in the Salton Trough, and a downstream-directed mode of river integration common in areas of crustal extension.

Keywords: Colorado River, Bouse Formation, Nevada, Arizona, flooding, Grand Canyon.

INTRODUCTION AND BACKGROUND

The development of the lower Colorado River is closely linked with incision of Grand Canyon, and both topics have been debated for most of the past century (e.g., Blackwelder, 1934; Longwell, 1936, 1947; Hunt, 1969; Lucchitta, 1972, 1979; Spencer and Patchett, 1997; Lucchitta et al., 2001). The Colorado River drains a large interior highland (including the Colorado Plateau and large parts of the southern and central Rocky Mountains) and then passes through the arid lowlands of the southern

House, P.K., Pearthree, P.A., and Perkins, M.E., 2008, Stratigraphic evidence for the role of lake spillover in the inception of the lower Colorado River in southern Nevada and western Arizona, *in* Reheis, M.C., Hershler, R., and Miller, D.M., eds., Late Cenozoic Drainage History of the Southwestern Great Basin and Lower Colorado River Region: Geologic and Biotic Perspectives: Geological Society of America Special Paper 439, p. 335–353, doi: 10.1130/2008.2439(15). For permission to copy, contact editing@geosociety.org.

Basin and Range Province before emptying into the Gulf of California (Fig. 1). The transition between the highland and lowland reaches is Grand Canyon, where the river is incised more than 1000 m below the surrounding plateau. Downstream from Grand Canyon, the Colorado River follows a peculiar course through the rugged terrain of the Basin and Range Province. The river's course initially tracks westward through a series of alternating alluvial basins and bedrock canyons in the Lake Mead area, and then it abruptly turns south through more canyons and basins before reaching the Gulf of California.

The course of the Colorado River downstream from western Grand Canyon clearly was established some time after 6 Ma (Lucchitta, 1979; Spencer et al., 2001), but the events that led to its development remain in some dispute (Spencer and Patchett, 1997; Lucchitta et al., 2001; Spencer and Pearthree, 2001; House et al., 2005a). Indeed, conflicting ideas about the development of the lower course of the Colorado River bear on the nature and timing of uplift of the Colorado Plateau and adjacent areas (Lucchitta, 1979; Spencer and Patchett, 1997; Lucchitta et al., 2001; Karlstrom et al., 2007). In this paper, we report on new geologic

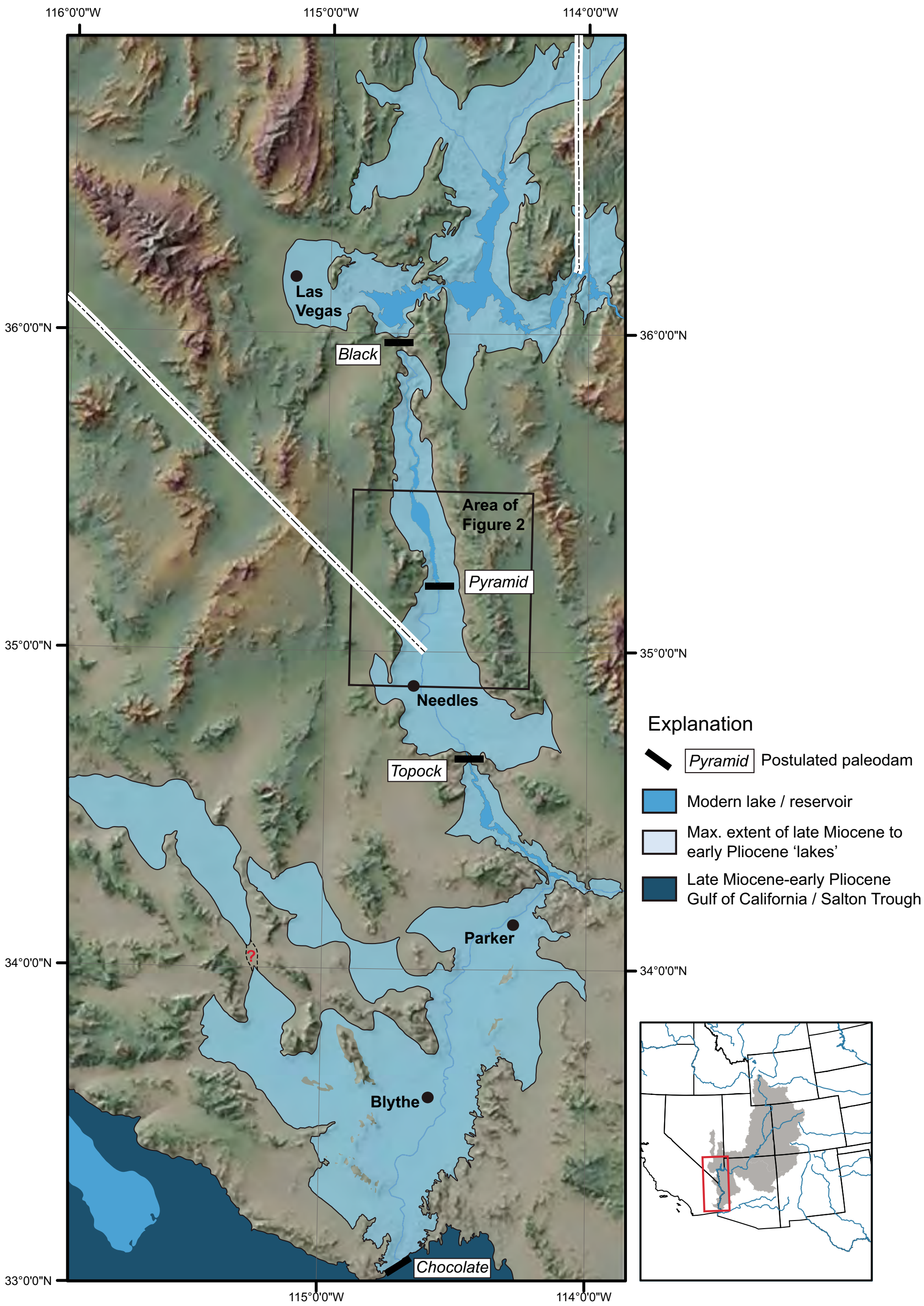


Figure 1. Overview of the lower Colorado River region from the western edge of Grand Canyon to the Salton Trough. Maximum extent of late Miocene lakes is indicated on basis of extrapolations of highest outcrops of Hualapai Limestone and Bouse Formation in relation to modern topography. Inset map shows the extent of the Colorado River Basin north of the confluence of the Colorado and Gila Rivers near Yuma, Arizona.

evidence from a central reach of the lower Colorado River that helps to clarify some aspects of the mechanism and timing of the river's inception.

Models for the Inception and Development of the Lower Colorado River

Three primary mechanisms have been proposed to explain the existence of and the particular features associated with the lower course of the Colorado River: (1) antecedence; (2) combinations of regional subsidence, marine incursion, headward erosion, drainage capture, and river progradation; and (3) downstream integration via lake spillover. The simplest explanation is antecedence, in which the river existed along its current course before uplift of the Colorado Plateau and adjacent areas and simply incised in response to the uplift (Powell, 1875; Dutton, 1882; Blackwelder, 1934). Abundant evidence that the lower Colorado River did not follow its present course before 6 Ma rules this possibility out, however (e.g., Lucchitta, 1979).

Fundamental differences in the latter two explanations stem primarily from alternate interpretations of the depositional environment of the Bouse Formation, a distinctive suite of sediments that occupy a key stratigraphic position between pre-integration and postintegration deposits in a string of alluvial basins between Hoover Dam and the Salton Trough (Fig. 1). The Bouse Formation includes a sequence of basal limestone and associated tufa overlain by mud, sand, and minor gravel of varying thicknesses (Metzger, 1968; Buising, 1990). It has been interpreted as a marine-estuarine deposit because it contains a limited assemblage of marine fossils in exposures as far north as Parker, Arizona (Metzger, 1968; Smith, 1970; McDougall, 2005). In contrast, strontium isotope ratios in Bouse carbonates throughout the extent of the deposit are more similar to modern Colorado River water than to seawater, which supports the hypothesis that Bouse sediments were deposited into lakes fed by the Colorado River (Spencer and Patchett, 1997; Poulson and John, 2003).

The marine-estuarine interpretation of the Bouse Formation supports an upstream-directed model of river integration involving headward erosion and drainage capture tied to regional subsidence (Lucchitta, 1972, 1979; Lucchitta et al., 2001). According to this model, a trough along the course of the Colorado River to near the mouth of Grand Canyon subsided to near or below sea level, allowing the sea to intrude far to the north from the Gulf of California. Base-level fall also drove incision and headward erosion into the western Grand Canyon area, eventually capturing the upper Colorado River and diverting it into the estuary. Sediment supplied by the Colorado River gradually filled in the estuary and forced the sea back to the south, possibly concurrent with regional uplift. In a purely marine interpretation, extant Bouse deposits 550 m above sea level near the northern limit of the unit require at least that much uplift since deposition.

A lacustrine interpretation of the Bouse Formation supports a downstream-directed integration model involving lake spillover through a series of closely spaced, enclosed basins. Blackwelder (1934) first proposed this mechanism for the development of the lower Colorado River. His hypothesis languished for decades but has recently been revitalized with new strontium isotope data linking Bouse basal carbonates throughout the system to water of the Colorado River (Spencer and Patchett, 1997). The current form of this hypothesis states that the Colorado River developed within some arrangement of downstream-spilling lakes beginning on the Colorado Plateau in the middle Tertiary (Meek and Douglass, 2001) and continuing through the area now occupied by Grand Canyon and the lower course of the Colorado River in the late Tertiary (Spencer and Patchett, 1997; Spencer and Pearthree, 2001; House et al., 2005a; Spencer et al., this volume). After the upper Colorado River integrated through the Grand Canyon area, it formed a series of lakes along the modern river's course. The river extended downstream as basins filled and successive bedrock divides were overtopped and lowered. The river reached the sea when a final divide north of Yuma was overtopped. This downstream-directed model does not depend on regional subsidence and subsequent uplift and only requires a sufficiently persistent influx of water to fill the basins and ultimately decreasing outlet elevations of each lake basin along the southward course of the river.

Critical evaluation of the competing models has been hindered by a lack of stratigraphic evidence documenting the geologic circumstances and timing of the arrival of the Colorado River into the region. In this paper, we report on new stratigraphic and tephrochronologic evidence from Mohave and Cottonwood Valleys (Fig. 2) that is consistent with the lake-spillover model. Our evidence comes from geologic mapping of late Miocene and early Pliocene sediments on both sides of an inferred bedrock paleodivide that would have separated the valleys. The stratigraphy of the deposits in each valley records a concurrent series of changes in depositional conditions that are consonant with short-lived lacustrine inundation in the upstream valley, flooding through the valley-separating divide, followed by an episode of deeper lacustrine inundation of both valleys. Tephrochronologic data indicate that these events occurred after 5.6 Ma. The deep lacustrine episode was followed by a period of erosion and then thick aggradation of sand and gravel associated with the Colorado River. Additional tephrochronologic data indicate that river aggradation culminated soon after 4.1 Ma and that net downcutting has dominated the history of the river since at least 3.3 Ma.

STUDY AREA

Cottonwood and Mohave Valleys are elongate, north-trending alluvial valleys along the lower Colorado River that lie roughly halfway between the mouth of Grand Canyon and the Gulf of California. The valleys are structural basins produced by major crustal extension in the middle Miocene (Howard and John, 1987; Spencer and Reynolds, 1989; Faulds et al., 1990). Our primary study area is approximately the southern half of Cottonwood Valley and the northern third of Mohave Valley. Light-colored Tertiary granitic rocks in the lower plate of a major detachment fault form the bulk of the Newberry Mountains on the

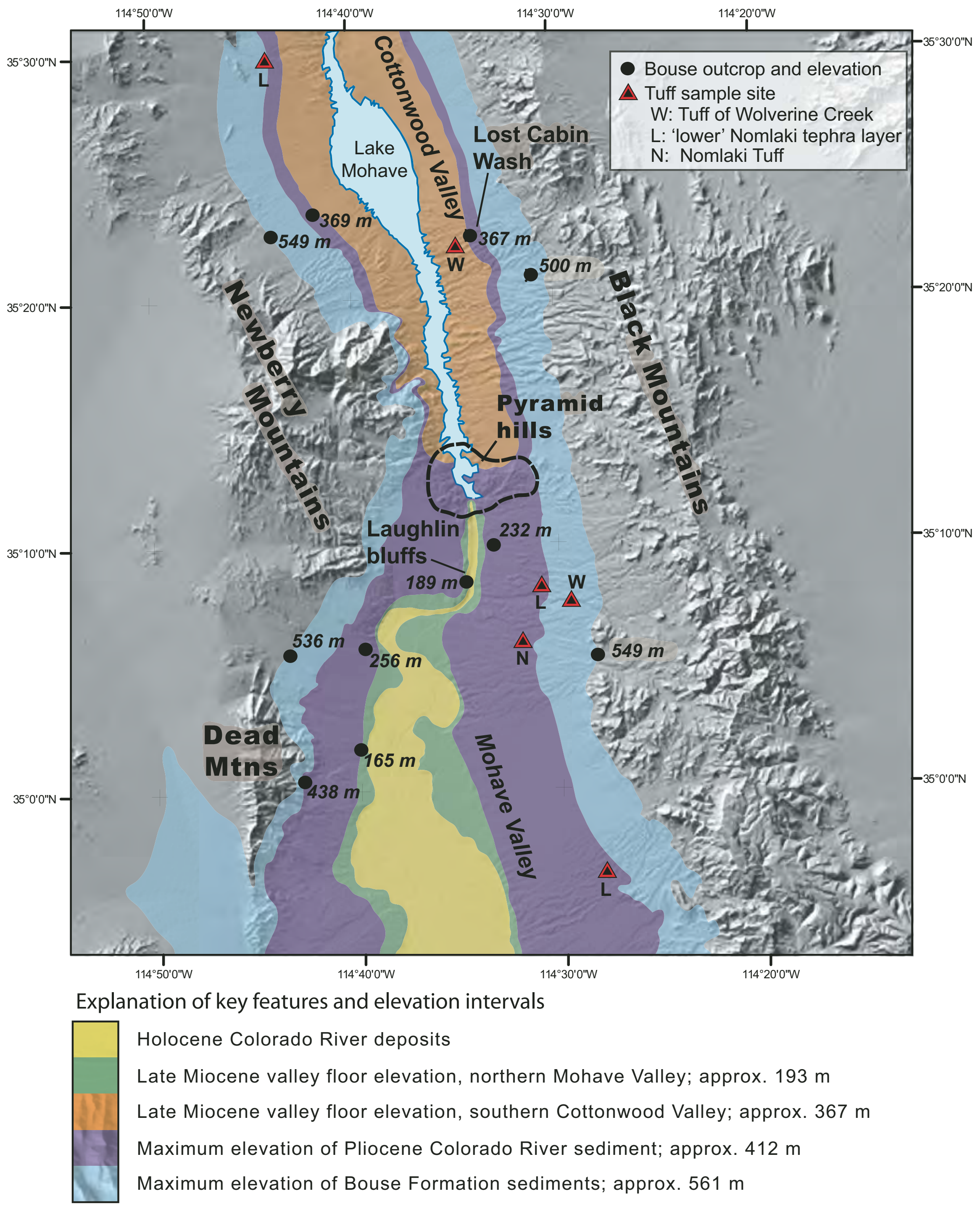


Figure 2. The study area in southern Cottonwood Valley and northern Mohave Valley showing tephra sample locations and selected Bouse Formation outcrops that indicate the vertical range of the unit in both valleys; corresponding elevations are noted in meters. Relief base is not a geologic map but is instead color-coded to indicate general geologic and topographic relations associated with pre–Bouse valley geometries, maximum Bouse water levels, and maximum extent of Pliocene Colorado River deposits.

west side of the study area, whereas upper-plate Tertiary volcanic and minor sedimentary rocks predominate in the Black Mountains on the east side. Alluvial deposits derived from each side of the valleys reflect this bedrock source dichotomy and help constrain preriver valley geometry. North of Cottonwood Valley, the Colorado River flows through Black Canyon, which separates this reach of the river from the west-trending reach in the Lake Mead area. The Pyramid hills (informal name) form the boundary between the Cottonwood and Mohave Valleys, and they are composed almost entirely of Proterozoic megacrystic granite (Faulds et al., 2004). The Colorado River has carved Pyramid Canyon through this divide. Topock Gorge, another rugged bedrock canyon, defines the southern margin of Mohave Valley. Substantial parts of southern Cottonwood Valley and northern Mohave Valley are deeply dissected (Fig. 2), and mapping and interpretation of sedimentary strata exposed in these areas provide most of the new information that we consider in this paper (Faulds et al., 2004; House et al., 2004, 2005b; Pearthree and House, 2005).

KEY STRATIGRAPHIC RELATIONS

Correlative deposits from each side of the Pyramid hills paleodivide document changes in depositional conditions associated with the arrival and early development of the Colorado River (Fig. 3). Important late Miocene to early Pliocene stratigraphic units exposed within ~20 km of the Pyramid hills paleodivide, from oldest to youngest, include: (1) postextensional alluvial-fan deposits derived from the valley-bounding mountains (Tfn and Tfb1; Newberry and Black Mountain fanglomerate); (2) axial valley facies, including fine gravel from the Black and Newberry Mountains interfingered with valley-filling silt and sand deposits (Tlcc and Tlcf; the Lost Cabin beds, found only in Cottonwood Valley); (3) gravelly alluvial fills in paleochannels cut in fanglomerate (Tng; Newberry channel gravel, found only in northern Mohave Valley); (4) coarse axial valley deposits dominated by clasts of Precambrian granite (Pyramid gravel, found only in northern Mohave Valley); (5) fine-grained deposits of the Bouse Formation, which typically include a thin basal limestone, locally grading upward into clay, silt, and sand beds (Tbl and Tbms; found in both valleys); (6) elaborately cross-stratified early Colorado River deposits of medium to coarse sand and gravel with abundant exotic, well-rounded clasts (Tcb; Bullhead alluvium, both valleys; generally correlative to unit B of Metzger and Loeltz, 1973); (7) local fan deposits interfingered with unit Tcb (Tfb2, both valleys); and (8) local fan deposits deposited on erosional surfaces cut across all older units (QTf, both valleys). We have identified three different tephra layers in the sequence that provide new temporal constraints on the development of the river in this reach. They include the 5.6 Ma tuff of Wolverine Creek; the 4.1 Ma lower Nomlaki tephra layer (informal name); and the 3.3 Ma Nomlaki Tuff. Detailed descriptions of these key units, their geochemical characteristics, and their age determinations are provided in Appendix 1 and Table A1.

The Cottonwood Valley Section

Late Miocene sedimentary strata in southern Cottonwood Valley (Fig. 3) suggest that deposition occurred in an enclosed basin with large alluvial-fan complexes extending to the valley axis from the Newberry and Black Mountains (units Tfn and Tfb1, Fig. 3). Exposures in the area of Lost Cabin Wash along the eastern shore of Lake Mohave reveal indurated, tilted fanglomerate deposits (Tft, Fig. 3) that date to the period of active normal faulting in the middle Miocene (Spencer and Reynolds, 1989). Gently eastward-dipping fanglomerate dominated by granitic clasts derived from the Newberry Mountains unconformably overlies the tilted deposits at ~240 m above sea level (a.s.l.). This Newberry fanglomerate grades into flat-lying axial valley gravel deposits that contain mixed clasts from the Newberry and Black Mountains (coarse-grained facies of the Lost Cabin beds, Tlcc). The axial gravel deposits grade upward into a sequence of flat-lying sandstone, siltstone, and mudstone beds with minor gravel (fine-grained facies of the Lost Cabin beds, Tlcf; Fig. 4A). The Lost Cabin beds grade laterally into local fanglomerate and contain several weak to moderately developed paleosols (Fig. 4B). We infer that the foregoing stratigraphic relations preclude the presence of a through-going Colorado River. Our interpretation is that the valley was an enclosed basin during the deposition of the Lost Cabin beds, and the axial drainage in southern Cottonwood Valley fed a depocenter to the north, the direction in which the valley widens and the fine-grained Lost Cabin beds thicken. The 5.59 ± 0.05 Ma tuff of Wolverine Creek (Fig. 4C; Table A1) is in the upper third of the fine Lost Cabin beds.

Fine Lost Cabin beds typically are overlain by Black Mountain fanglomerate (Tfb2 and upper part of Tfb1) along an erosional unconformity at an elevation of 350 m a.s.l. or less. In some locations, however, thin beds of Bouse limestone and calcareous mud interfinger with the upper few meters of the Lost Cabin beds (Fig. 5A). Thick and extensive Bouse limestone, sand, and mud deposits overlie a minor, locally erosional unconformity above this key interval (Fig. 5B). In places, the unconformable contact is characterized by small channels filled with local gravel and reworked mud (Fig. 5C). This stratigraphic relationship suggests a phase of intermittent lacustrine and subaerial sedimentation separated from a more extensive and prolonged period of lacustrine deposition by a period of erosion. At higher elevations to the east, Bouse limestone rests on gently west-dipping fan paleosurfaces underlain by weathered Black Mountain fanglomerate. In a few localities, up to 10 m of mudstone, siltstone, and sandstone are preserved above the basal limestone (Fig. 5D). We have found Bouse limestone, tufa, and related clastic shoreline sediments at elevations up to 550 m a.s.l. in central Cottonwood Valley, indicating a local water depth of at least 200 m (Fig. 2). The upper Bouse contact typically is erosional, and Bouse deposits and some underlying Lost Cabin beds were removed prior to renewed fanglomerate deposition.

Early Colorado River sand and gravel deposits (Bullhead alluvium), consisting of a mix of well-rounded, exotic gravel and subangular local gravel, rest on an erosional unconformity

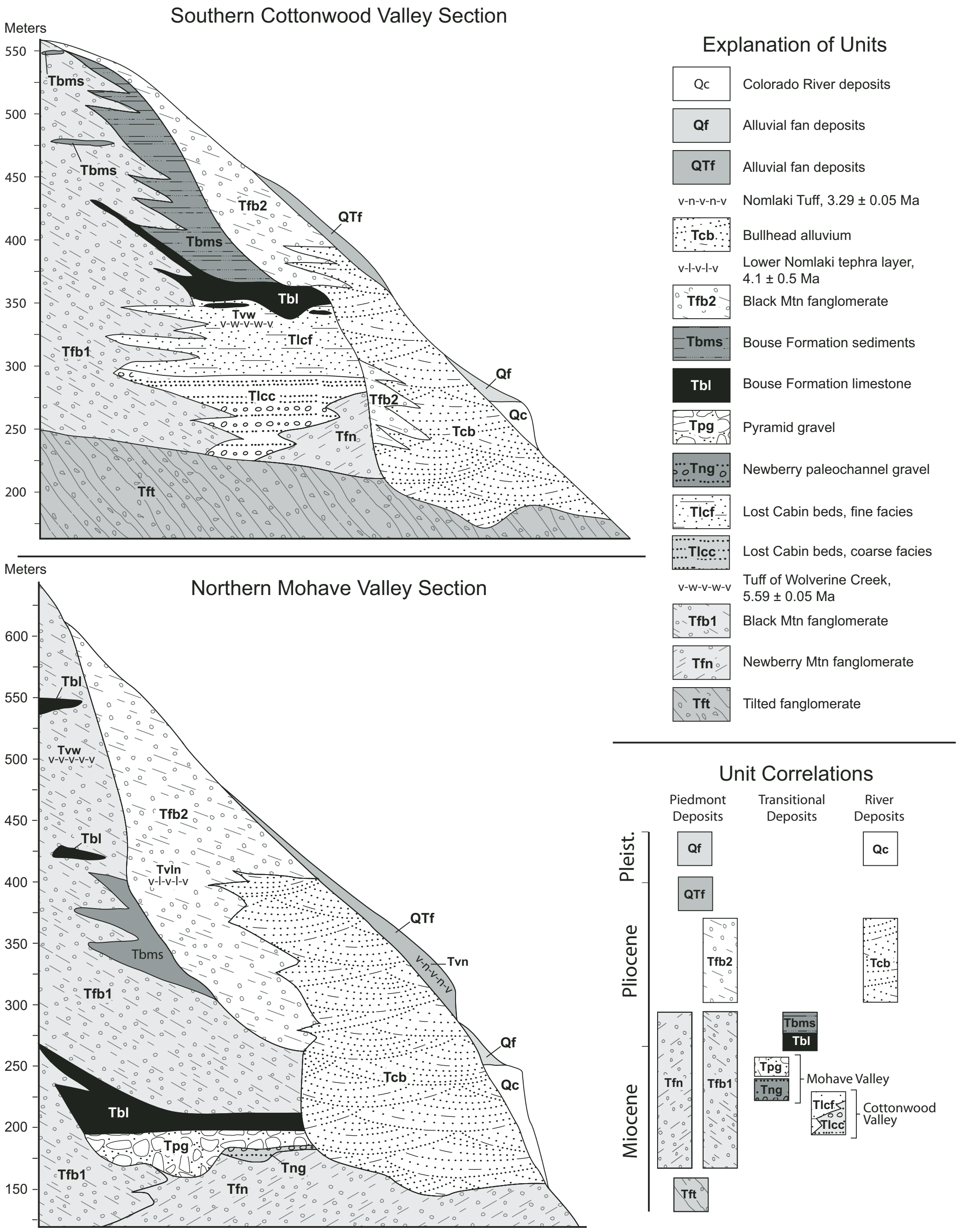


Figure 3. Schematic diagrams of key late Cenozoic stratigraphic relations in Cottonwood and Mohave Valleys. Vertical exaggeration ~10×. Upper section: Southern Cottonwood Valley. Note that the contact between the Lost Cabin beds (Tlc) and the Bouse limestone (Tbl) indicates the approximate valley axis elevation prior to river integration. Lower section: Northern Mohave Valley. Note that the contact between the Newberry paleochannel gravel and the Pyramid gravel indicates the approximate valley axis elevation prior to river integration. Unit explanations and correlation diagram on right side of figure apply to both sections. Transitional units exclusive to each valley are indicated on correlation diagram.

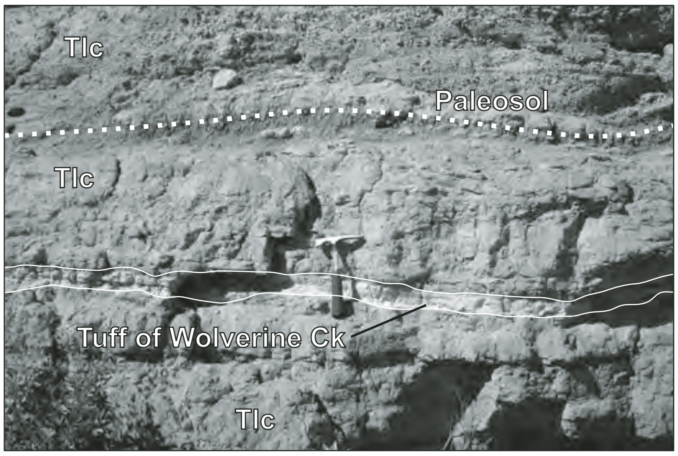


Figure 4. Photographs of the Lost Cabin beds in the Lost Cabin Wash area. (A) Thick section of fine-grained Lost Cabin beds (Tlc) overlain by Black Mountain fanglomerate (Tfb). (B) Paleosol in the Lost Cabin beds overlying the tuff of Wolverine Creek. Note rock hammer for scale. Site is in Odyssey wash (informal name) in vicinity of Lost Cabin Wash.

carved deeply into all of the older deposits. In southern Cottonwood Valley, the lowest exposed outcrops of the Bullhead alluvium are at the level of Lake Mohave (~195 m a.s.l.), and they have been found as high as 427 m in the valley. Gently west-dipping Black Mountain gravel deposits (QTf) with massive duripans up to 3 m thick cap the highest ridges on the piedmont. These piedmont deposits rest unconformably on older Black Mountain fanglomerate (Tfb) and Bullhead alluvium and were presumably graded to a late Pliocene paleoriver level of ~300–330 m a.s.l.

The Northern Mohave Valley Section

Stratigraphic relations in northern Mohave Valley (Fig. 3) also record a late Miocene to early Pliocene transition from local drainage to deep inundation to through-going Colorado River, and they provide an interesting complement to the Cottonwood Valley sequence. Conspicuous differences between the late Miocene to early Pliocene strata in each valley exist and are interpreted to have resulted from their locations relative to the inferred paleodivide between the valleys. The lowest strata exposed in northern Mohave Valley consists of local fanglomerate, which at various locations contains gravel derived from the Newberry Mountains, the Black Mountains, or the Pyramid hills (Faulds et al., 2004). This is consistent with alluvial-fan deposition from the western, eastern, and northern flanks of the valley and drainage to a depocenter to the south.

A particularly illustrative example of the northern Mohave Valley section is exposed in the bluffs astride the Colorado River south of Laughlin, Nevada (Figs. 2 and 6A). There, a deposit of cross-stratified, locally derived gravel (Newberry paleochannel gravel) fills relatively small, roughly south-trending paleochannels cut into Newberry fanglomerate (Fig. 6B). The fanglomerate and paleochannels are overlain along an erosional unconformity by a distinctive coarse-grained fluvial conglomerate, the composition of which is dominated by clasts of dark-colored megacrystic granite and lesser amounts of gravel eroded from local fanglomerate deposits (Fig. 6C). We have not identified any diagnostic Colorado River sediments in this gravel unit. The nearest source of the dominant lithology (megacrystic granite) is in the Pyramid hills paleodivide ~7 km to the north, and we have informally named the unit the Pyramid gravel. The Pyramid gravel is a broadly tabular deposit up to 30 m thick, is crudely to moderately stratified, and has clast-supported and matrix-supported beds. Clast imbrication and trough cross-stratification are evident in several channel-filling exposures. Boulders up to 1 m in intermediate-axis diameter are common in the lower part of the deposit, but it typically is a pebble and cobble conglomerate. The basal limestone of the Bouse Formation overlies the Pyramid gravel in the Laughlin bluff section along a sharp, relatively flat contact. This stratigraphic pairing indicates an abrupt change from high- to low-energy depositional conditions (Fig. 6D). The entire foregoing sequence is overlain along a high-relief erosional unconformity by the Bullhead alluvium (Fig. 6E).

At other locales in Mohave Valley, basal Bouse deposits overlie paleo–alluvial-fan surfaces and bedrock slopes and are interbedded with fanglomerate deposits up to elevations of 550 m a.s.l. (Fig. 7). High on the east side of the valley, the tuff of Wolverine Creek (Table A1) is interbedded with fanglomerate deposits 11 m below the Bouse limestone, providing a maximum age constraint of 5.6 Ma for deep inundation of Mohave Valley (Fig. 7). The basal limestone beds are everywhere less than a few meters thick but are locally overlain by up to 30 m of mud and sand. The distribution of outcrops of the Bouse Formation suggests that water depth in northern Mohave Valley may have exceeded 400 m, and the general form of the valley at that time was similar to its present form (Fig. 8).

As in Cottonwood Valley, a thick deposit of Bullhead alluvium rests on an unconformity that cuts across all older basin deposits in Mohave Valley. The lowest exposures of Bullhead alluvium in both valleys contain abundant locally derived gravel mixed with well-rounded quartzite, chert, and other exotic

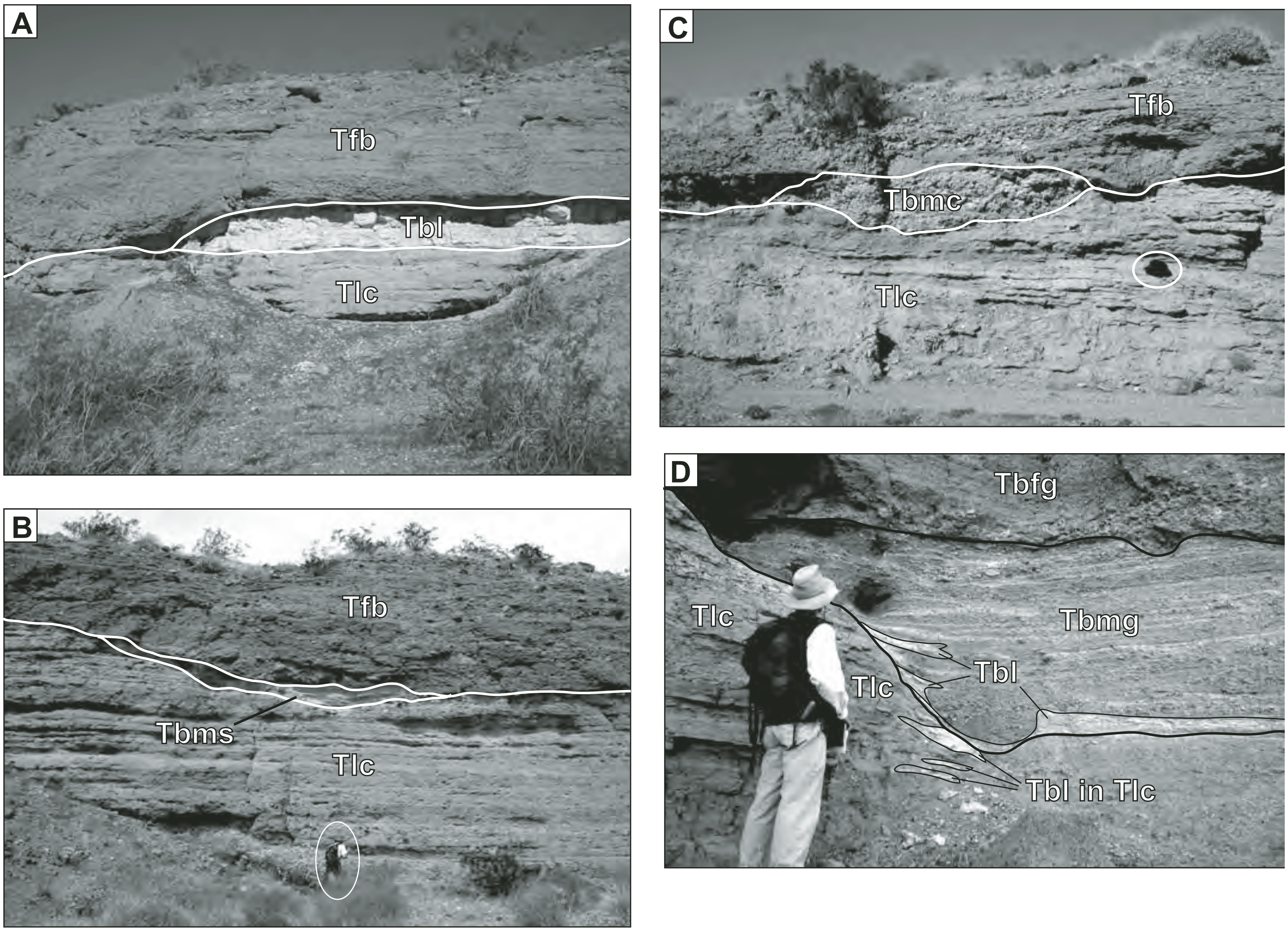
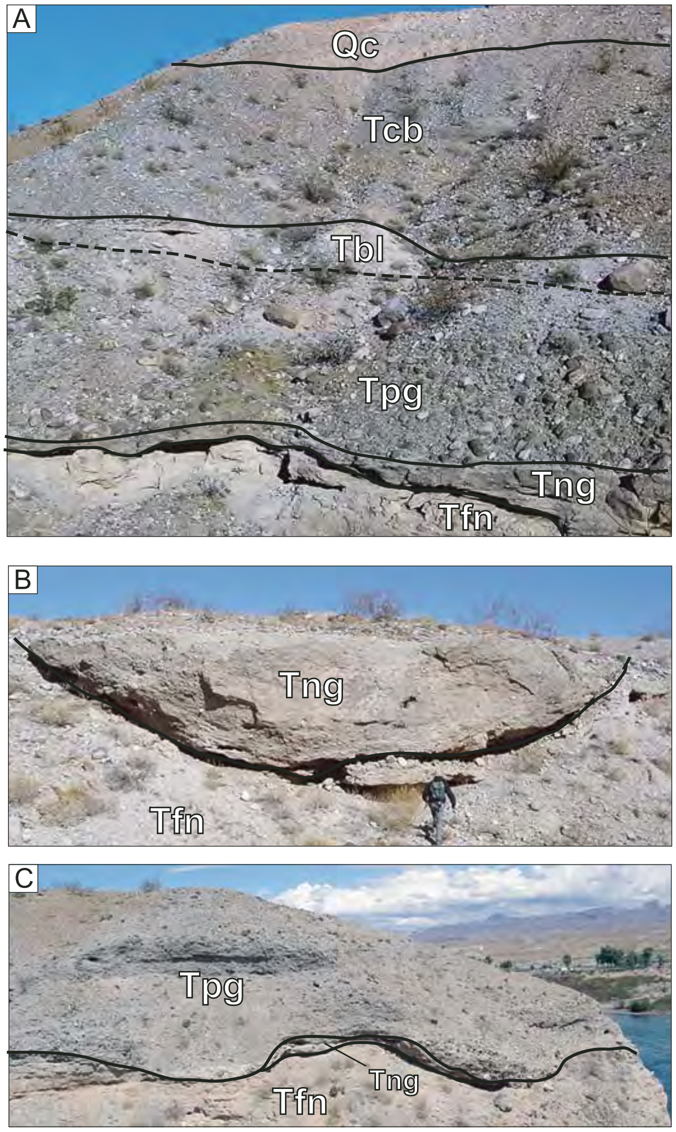


Figure 5. Stratigraphic relations between the Lost Cabin beds and the Bouse Formation. (A) Coarse Lost Cabin beds (Tlc) overlain by Bouse limestone (Tbl), which is in turn unconformably overlain by Black Mountain fanglomerate (Tfb). Thickness of Tlc in photo ~1 m. (B) Paleochannel at top of Lost Cabin beds (Tlc) filled with calcareous Bouse mud and sand (Tbms). Note person for scale. (C) Paleochannel in Lost Cabin beds (Tlc) filled with pebbly Bouse mud-ball conglomerate (Tbmc). Note hat (circled) for scale. (D) Contact between Lost Cabin beds (Tlc) with interfingered marl (Tbl) overlain along erosional unconformity by muddy gravel (Tbmg) with interbedded marl (Tbl). Disconformable upper beds are possible wave-worked fanglomerate (Tbfg).

pebbles and small cobbles (Figs. 9A and 9B). Exposures of Bullhead alluvium higher in the section on the Black Mountain piedmont in Mohave and Cottonwood Valleys contain laterally extensive, tabular beds of trough cross-stratified gravel, thick beds of complexly cross-stratified sand, and minor flat-lying mud. Overall, the Bullhead alluvium is a very complex fluvial deposit with numerous stratigraphic discontinuities (e.g., Fig. 12 *in* Metzger and Loeltz, 1973); however, to date we have not recognized any paleosols or erosion surfaces suggestive of major hiatuses or cut-and-fill episodes in our study area. Bullhead alluvium is extensively interstratified with piedmont fanglomerate deposits, particularly in the upper part of the section. Two outcrops of the 4.1 ± 0.5 Ma lower Nomlaki tephra layer (Table A1) rest in fine tributary fan gravel at elevations of 365–390 m a.s.l., but Bullhead alluvium is found at stratigraphically higher positions up to 400 m a.s.l. in the immediate vicinity of each tephra layer exposure in Mohave Valley (Fig. 10). Thus, the lower Nomlaki tephra is very near the top of the Bullhead aggradational sequence.

Relatively thin, west-dipping Black Mountain piedmont gravels with very strongly developed duripans (Herriman and Hendricks, 1984) lie above an erosion surface cut on Bullhead alluvium and Black Mountain fanglomerate. The 3.3 Ma Nomlaki Tuff (Table A1) is intercalated in these piedmont deposits at elevations of 395 and 350 m a.s.l., so they must have been graded to a former, post-Bullhead level of the Colorado River below 350 m a.s.l. In summary, the tephrochronologic and stratigraphic evidence support an interpretation that the Bullhead aggradation event in the study area culminated sometime around ca. 4.1 Ma, and incision into the deposit of at least 50 m had occurred by ca. 3.3 Ma.



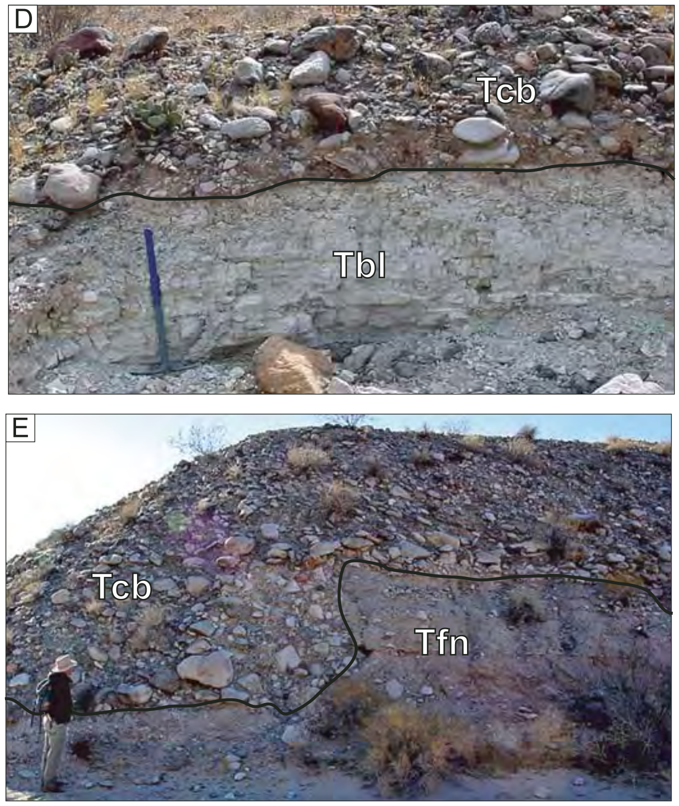


Figure 6. Photographs of the Laughlin Bluffs section. (A) Complete section (~26 m thick) showing stratigraphic relation between each key unit (Tfn—Newberry Mountain fanglomerate; Tng—Newberry paleochannel gravel; Tpg—Pyramid gravel; Tbl—Bouse limestone; Tcb—Bullhead alluvium; Qc—Younger Colorado River sediments). (B) Newberry gravel (Tng) filling paleochannel in fanglomerate (Tfn). (C) Pyramid gravel flood deposit (Tpg) filling paleochannels in Newberry fanglomerate (Tfn). (D) Bouse marl (Tbl) disconformably overlain by Bullhead alluvium (Tcb). (E) Base of the Bullhead alluvium incised into Newberry fanglomerate.

THE DEVELOPMENT OF THE COLORADO RIVER IN COTTONWOOD AND MOHAVE VALLEYS

Any proposed scenario for the development of the Colorado River in this region must accommodate several key stratigraphic relationships and temporal constraints in Mohave and Cottonwood Valleys. No through-going drainage connected the valleys before 5.6 Ma. In Cottonwood Valley, an interval of fine-grained clastic deposition by local streams was ongoing by 5.6 Ma and was terminated with an erosional interval followed by limestone deposition. There is strong evidence for a southward-directed drainage divide failure between the valleys before deep inundation of both valleys. A period of erosion followed the deep inundation and preceded the arrival of voluminous coarse Colorado River sediment. After its arrival, the river aggraded dramatically until shortly after 4.1 Ma. River incision had reached well below its level of maximum aggradation by 3.3 Ma and has remained below that level to the present. The entire series of transitional events from enclosed basins to deep-water inundation followed by the arrival of the Colorado River and thick aggradation, and finally the initial river incision appears to have transpired in less than 2 m.y.

We propose the following scenario for the early development of the Colorado River based on interpretation of the evidence that we have compiled from this area (principal events summarized in Fig. 11):

1. Through much of the late Miocene, the Cottonwood and Mohave Valleys had separate, closed drainage systems (Fig. 11A) that were relicts of middle Miocene extensional faulting as shown by local fanglomerates, axial valley gravel deposits, and inferred playa deposits north and south of the primary study area.

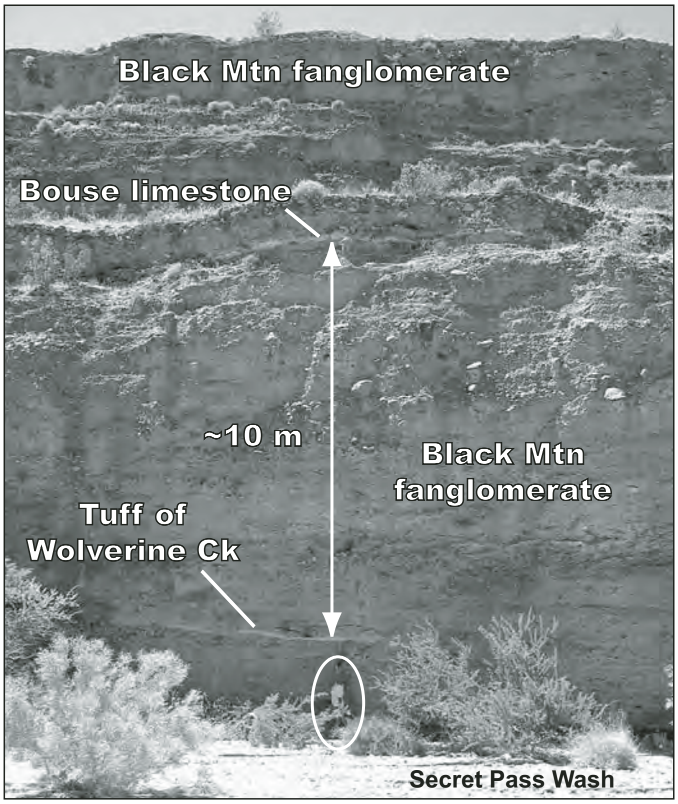


Figure 7. Photograph of the tuff of Wolverine Creek and overlying, thin bed of Bouse limestone in Secret Pass Wash (modified from House et al., 2005a). Person indicated in lower part of photo is for scale.

2. Exotic water and sediment, possibly associated with the Colorado River, began to enter Cottonwood Valley from the north in the late Miocene (Fig. 11B). This resulted in the expansion of fine-grained, primarily subaerial deposition into southern Cottonwood Valley from the north. This interval of deposition was ongoing by 5.6 Ma, as shown and dated by the Lost Cabin beds and the tuff of Wolverine Creek.

3. The influx of water eventually formed a lake in Cottonwood Valley (Fig. 11C), represented by thin beds of limestone interfingered with upper Lost Cabin beds.

4. Water began to flow over or through extensively fractured bedrock in the Pyramid hills, as indicated by the Newberry paleochannel gravel. Eventually, the divide was catastrophically breached, as shown by the Pyramid gravel (Fig. 11C). In Cottonwood Valley, this event is recorded as an erosional interval separating the Lost Cabin beds and the bulk of the Bouse Formation.

5. Persistently inflowing Colorado River water was blocked, possibly at the south end of Mohave Valley, resulting in deep inundation of both Mohave and Cottonwood Valleys (Fig. 11D). The maximum water surface elevation in both valleys was ~550 m a.s.l. as inferred from the highest outcrops of Bouse limestone, sandstone, and tufa in both valleys. During this time, concurrent influx of fine sediment and reworking of local sediments resulted in deposition of clastic sediments of the Bouse Formation, possibly in the form of a delta extending from northern Cottonwood Valley.

6. The paleodivide to the south was ultimately overtopped and eroded in some manner, resulting in lake drainage and erosion of Bouse and older deposits (Fig. 11E).

7. Voluminous Colorado River bed-load sediment arrived into subaerially exposed valleys (Fig. 11E) and was first deposited with abundant locally derived sediments at levels near the modern river

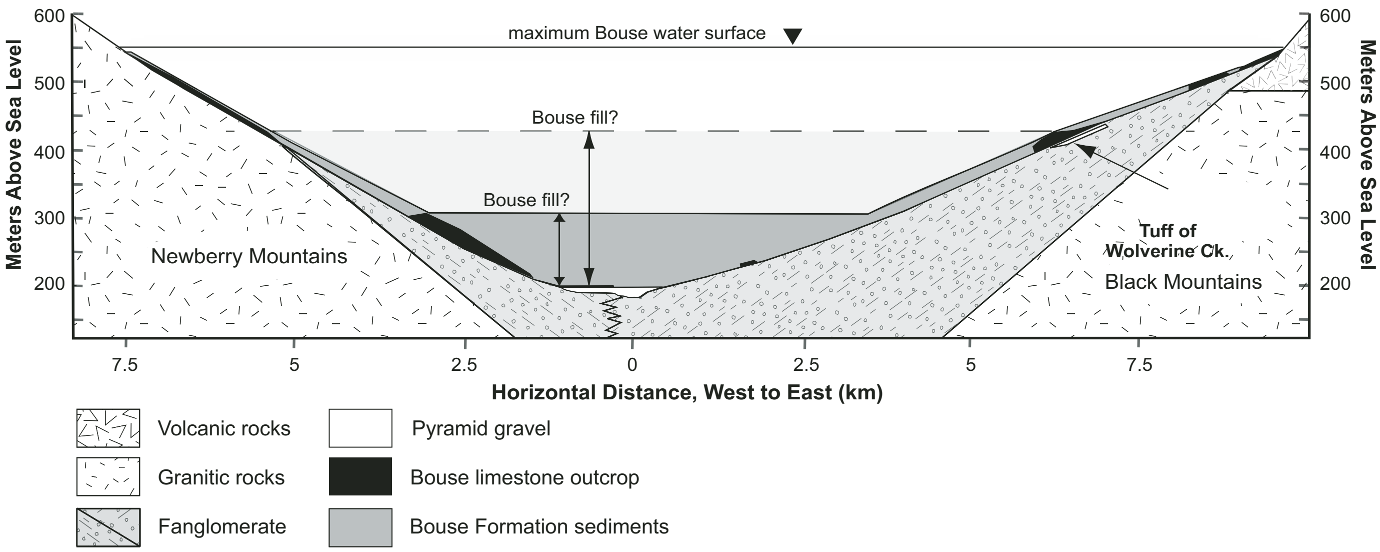


Figure 8. Schematic diagram showing distribution of Bouse limestone, tufa, and clastic sediments in Mohave Valley and speculations on extent of Bouse fill. Extant outcrops of Bouse Formation are shown in black (adapted from House et al., 2005a).

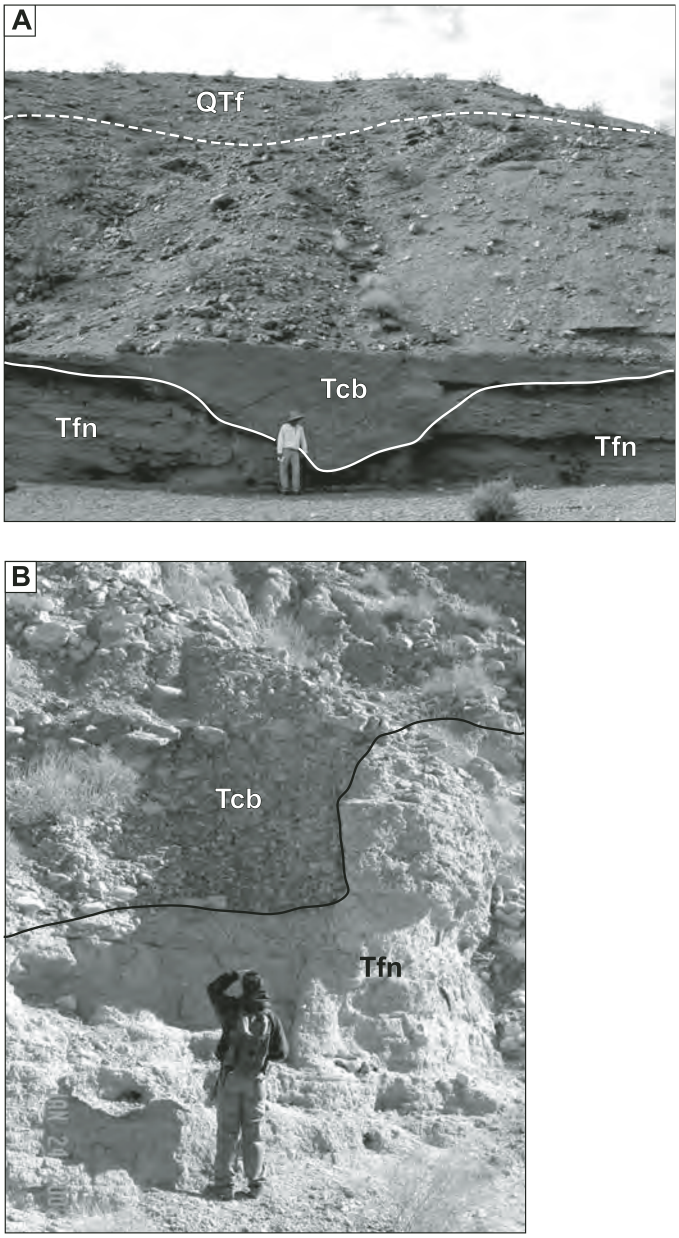


Figure 9. Examples of the lower, erosive contact at the base of Bullhead alluvium in Cottonwood and Mohave Valleys. (A) Paleochannel in Granite Wash, Cottonwood Valley, ~25 m above Lake Mohave surface (195 m at mouth of wash). Deposit is poorly sorted, massive to cross-stratified pebbly conglomerate with mix of dominantly local gravel and sparse, exotic, well-rounded pea gravel. Bullhead paleochannel is cut into east-dipping Newberry Mountain fanglomerate and is overlain uncomformably by younger alluvial fan deposits (QTf). (B) Base of gravelly Bullhead alluvium incised in Newberry fanglomerate in the Laughlin bluffs area (Panda gulch of House et al., 2005a). Note conspicuous contact. Deposit is mix of locally derived coarse, subrounded gravel and exotic, well-rounded pea gravel and cobbles. Lowest exposure of Bullhead alluvium in Panda gulch is 18 m above modern Colorado River level (152 m at mouth of gulch). Tfn—fanglomerate; Tcb—Bullhead alluvium.

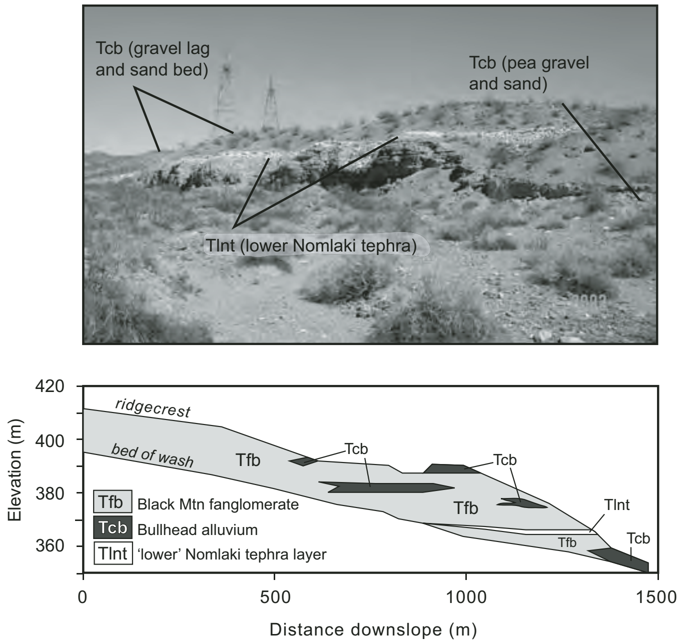


Figure 10. The "lower" Nomlaki tephra layer interfingered with Black Mountain fanglomerate and Bullhead alluvium, Powerline outcrop, Mohave Valley, Arizona. Upper image is photo of tephra layer outcrop in fanglomerate with strata of Bullhead sediment and surface gravel lags indicated; lower image is a cross section of the same site showing interpreted stratigraphic relations.

~150–200 m a.s.l. This is indicated by the lower part of the Bullhead alluvium and its lowest contact with underlying fanglomerates.

8. Massive, presumably integration-driven aggradation of the Colorado River and tributary fanglomerates filled Mohave and Cottonwood Valleys with sediment to a level of ~400 m a.s.l. (Fig. 11F). This aggradational interval culminated shortly after 4.1 Ma, as indicated by stratigraphic relations among the Bullhead alluvium, local fanglomerate, and the lower Nomlaki tephra layer.

9. Incision of the Colorado River below the maximum level of aggradation began before 3.3 Ma (Black Mountain piedmont gravel, Nomlaki Tuff) and ultimately continued to a level near modern river grade, resulting in the removal of vast amounts of Bullhead alluvium and local fanglomerate (not specifically shown on Fig. 11, but similar to Fig. 11E).

DISCUSSION: REGIONAL GEOLOGIC CONNECTIONS AND PROCESS IMPLICATIONS

The postulated sequence of events outlined here is consistent with recent geochemical studies that have tied the development of the lower Colorado River to lacustrine deposits elsewhere along the river's course (Spencer and Patchett, 1997; Poulson and John, 2003). It is also consistent with stratigraphic and geochronologic evidence for development of the Colorado River in areas upstream (e.g., Howard and Bohannon, 2001) and downstream (e.g., Buising, 1990; Dorsey et al., 2007) as described next.

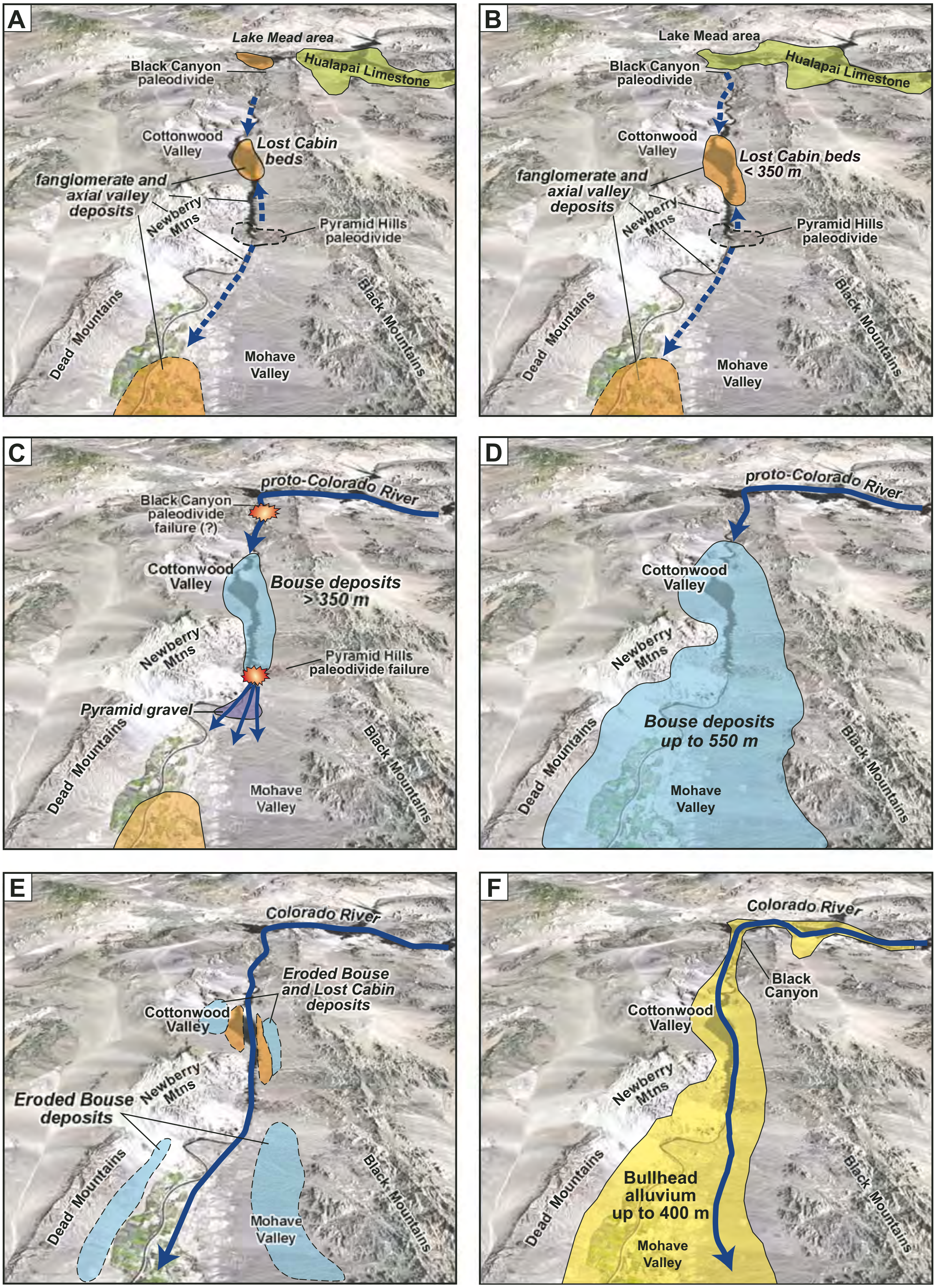


Figure 11. Proposed stages in the evolution of the lower Colorado River based on evidence and interpretations presented in this paper. Base map was provided courtesy of William A. Bowen. See text for discussion of each panel.

The Muddy Creek Formation and the Hualapai Limestone

The basic stratigraphic associations described in Cottonwood and Mohave Valleys are similar to relationships upstream in the Lake Mead area, where longer-lived lacustrine conditions immediately preceded the arrival of the Colorado River. A typical pre–Colorado River stratigraphic package found in basins in this area includes the Muddy Creek Formation (Longwell, 1936; Bohannon, 1984), which consists of a thick sequence of primarily subaerial, siliciclastic and evaporite deposits that accumulated between ca. 15 and 6 Ma. The Hualapai Limestone is a lacustrine carbonate unit that is interbedded with or overlies siliciclastic or evaporite deposits in the Muddy Creek Formation along the course of the Colorado River in the Lake Mead area. Limestone deposition occurred between ca. 11 and 6 Ma (Lucchitta, 1979; Faulds et al., 2001; Spencer et al., 2001). In the Grand Wash Trough, at the mouth of the Grand Canyon, limestone beds interfinger with clastic deposits lower in the section and become cleaner and more extensive higher in the section (Lucchitta, 1979). Excluding beds that have been deformed by movement on the Wheeler fault zone, the uppermost Hualapai Limestone beds in the Grand Wash Trough are ~900 m a.s.l. (Howard and Bohannon, 2001). Farther west in the Lake Mead area, thinner Hualapai deposits overlie a varied landscape (Lucchitta, 1979), and the highest outcrops are ~700 m a.s.l. (Howard and Bohannon, 2001). A similar package of limestone overlying siliciclastic sediments, including the 5.6 Ma tuff of Wolverine Creek and evaporites, is exposed in a small basin northeast of Las Vegas (Castor and Faulds, 2001; Castor et al., 2000). The highest limestone deposits there are ~670 m a.s.l.

We postulate that an influx of Colorado River water through the developing Grand Canyon resulted in deposition of at least the upper part of the Hualapai Limestone, deepening and expanding preexisting lakes (Figs. 11A and 11B). The basins in the western Lake Mead area were ultimately inundated by a sizable lake that filled to >700 m a.s.l. and then spilled through the Black Canyon area and initiated or accelerated the filling of Cottonwood Valley with sediment and, ultimately, water (Fig. 11A). The lacustrine deposits in the Lake Mead area may have been locally overlapped by Colorado River sediment, but in most places even the highest Colorado River alluvium is inset below the highest level of Hualapai deposition (Howard and Bohannon, 2001). This implies that the lake basins had drained and had undergone some erosion before the arrival of substantial amounts of Colorado River bed load.

Bouse Associations between Laughlin, Nevada, and Parker, Arizona

Deposits of the Bouse Formation are relatively extensive along the lower Colorado River downstream from Mohave Valley, and they have stratigraphic and geographic characteristics that can be reconciled with a downstream-directed lake-spillover model of integration. For example, in the Parker, Arizona, area, Buising (1988, 1990) described a stratigraphic relation in which a distinctive cross-stratified deposit of fluvial gravel that is interfingered with the base of the Bouse Formation overlies locally derived fanglomerate. The cross-stratified gravels contain far-traveled sediment and have sedimentary structures indicative of generally southward transport (Buising, 1990). These relations suggest that an influx of water from the north preceded and accompanied basal Bouse deposition in this area. Buising (1990) interpreted this association as evidence for a fluvial-marine interface; however, it is also consistent with a fluvial-lacustrine interface similar to, but more well developed than, what we have described in the Laughlin area.

If the Bouse Formation at Parker records an interface between the early Colorado River and the sea, then basal Bouse and interfingered Colorado River deposits there would have to be older than deposits in the Laughlin area. Additionally, there should be some evidence for northward transgression of the fluvial-marine interface through the intervening areas. However, at Laughlin, the (younger in the case of transgression) Bouse Formation overlies a south-directed divide-breach deposit composed of locally derived sediments. The only Colorado River deposit in the northern Mohave Valley section overlies the Bouse limestone along a major disconformity. There is no evidence of a through-flowing Colorado River below the lowest Bouse outcrops at Laughlin. This argues against the presence of the river in the Parker area before the divide between Mohave Valley and Cottonwood Valley was breached.

The simplest, but not necessarily the only, explanation of the Bouse stratigraphy in each location is that drainage from a lake in Mohave and Cottonwood Valleys delivered southward-flowing water into the Parker area, ultimately forming a large lake into which the Bouse Formation and early Colorado River sediment were deposited.

Distribution and Elevations of the Highest Bouse Outcrops

The elevations of highest Bouse limestone outcrops along the lower Colorado River form much of the foundation for the marine transgression–rapid uplift model. However, the profile of outcrop elevations has an upward-stepped progression from south to north, and the steps coincide with locations of likely bedrock divides between successive basins (Spencer et al., this volume). For example, the highest Bouse Formation outcrops from the Lake Havasu area in the north to the Chocolate Mountains in the south are consistently 330 m a.s.l., ~220 m lower than in Mohave Valley (Spencer et al., this volume). This dramatic decrease in water-surface elevation is consistent with separate lake basins and is difficult to reconcile with a continuous marine incursion. In our study area, the highest outcrops of Bouse marl, tufa, and clastic sediments that we have found are all at similar elevations along ~33 km of the axis of the valleys (Fig. 2) and are interpreted as shoreline deposits that record inundation of both valleys by the same deep lake. It is reasonable to suspect that had significant regional uplift occurred, the highest extant shoreline deposits separated by the greatest north-south distance would be at distinctly different elevations.

Directional Age Controls on the Bouse Formation

Determination of the age of the Bouse limestone in each basin would provide important clues about the deposit's origin. In a marine transgression scenario, one would expect that the limestone would "young" to the north, whereas in a lacustrine scenario, the opposite would be the case as lakes spilled downstream. Presently available data about directional age trends in the Bouse Formation are equivocal, however. Our tephrochronologic data constrain the age of the Bouse in Mohave and Cottonwood Valleys to between 5.6 and 4.1 Ma. We suspect, but do not know, that the age of the Bouse is closer to 5.6 than to 4.1 Ma, since the latter age corresponds to the near culmination of a series of major events that occurred after Bouse deposition in our study area. Downstream from our field area, the ca. 4.8 Ma Lawlor Tuff (Sarna-Wojcicki et al., 1991) sits near the top of Bouse limestone deposits in the Parker-Blythe area. In combination, these constraints only provide weak support for decreasing age of Bouse deposition in a downstream direction.

Fossils in the Bouse Formation

The limited salt-water faunal assemblages identified in the Bouse Formation in the Parker-Blythe basin (Smith, 1970; Lucchitta et al., 2001; McDougall, 2005) are the principal remaining support for a marine incursion. The lowermost Bouse lake would have encompassed an immense area extending from Parker Valley to the Chocolate Mountains and westward into a series of low-lying basins in the Mohave Desert (Fig. 1). Recent hydrologic modeling suggests that it may have taken tens of thousands of years to fill this extensive lake to overflowing due to likely high rates of evaporation; consequently, the lake water could have become quite saline prior to spilling into the Yuma area (Spencer et al., this volume). This might have allowed salt-water fauna to survive in the lake, but the mechanism for transportation of marine fauna into such a lake, if it existed, is disputed (Lucchitta et al., 2001; Spencer and Patchett, 1997).

Distribution of Pliocene Colorado River Deposits

An increasing body of evidence indicates that an integrated Colorado River was traversing the Colorado Plateau and building a delta near the latitude of its modern terminus by the early Pliocene. Evidence from near the mouth of the Grand Canyon indicates that that the earliest integrated Colorado River flowed on top of the Hualapai Limestone, and the river had incised into the limestone before 4.7 Ma (Howard and Bohannon, 2001). Our studies indicate that the river was approaching its maximum level of aggradation in Mohave and Cottonwood Valleys by ca. 4 Ma. Paleobotanical evidence from Colorado River deposits near Yuma indicates that high-standing river gravels were deposited there in approximately the middle Pliocene (Nations et al., 1998). Recent paleomagnetic and paleontologic studies of sediments in the Salton Trough suggest that the first arrival of Colorado River sand could possibly have occurred as early as 5.33 Ma, the Miocene-Pliocene boundary (Dorsey et al., 2007). This age, however, is difficult to reconcile with the Lawlor Tuff constraint mentioned previously, because its presence in the Bouse would preclude the arrival of the river into the Salton Trough prior to ca. 4.8 Ma, barring a particularly complex series of events.

The timing of the first arrival of Colorado River alluvium in the Salton Trough suggests that the entire south-trending part of the lower Colorado River developed after 5.6 Ma (the age of the tuff of Wolverine Creek) and possibly as early as 5.33 Ma (Dorsey et al., 2007) or maybe after 4.8 Ma (age of Lawlor Tuff). In either case, this geologically short interval of Bouse deposition and river development is more consistent with relatively rapid filling and spilling of a series of lakes than with a sequence of regional subsidence and marine transgression, immediately followed by regional uplift and concurrent, thick river aggradation. Furthermore, the presence of a through-going river below the edge of the Colorado Plateau before 4.7 Ma and the accumulation of a thick alluvial fill in valleys downstream by ca. 4.1 Ma strongly suggest a linkage among erosional and depositional events that is neither dependent on nor fully consistent with regional uplift along the river's entire lower course. The simplest explanation is that the episode of massive early Pliocene aggradation along the lower Colorado River was forced by upstream integration and canyon incision, and it was minimally influenced by tectonic events.

River and Lake Associations in the Western United States

Linkages among rivers and transient lakes are common in the geologic record, particularly in areas of crustal extension (Potter, 1978; Cohen, 2003). This circumstance prevails in the Basin and Range Province of the western United States where Miocene extension produced an array of closely spaced basins bounded in many cases by high-standing, relatively well-watered mountain ranges. In this region, there are many examples where runoff from persistent, orographically enhanced precipitation in large highland areas fed rivers that formed lakes in arid to hyperarid, low-lying basins. Time-varying hydrologic inputs and regional topographic constraints have resulted in different extents of basin interconnection as valley divides have been intermittently or permanently overtopped.

Several examples of rivers in the Great Basin illustrate different degrees of fluvial integration related to filling and spilling of series of enclosed basins. Throughout the Pleistocene, fluctuating hydrologic inputs to the Lahontan Basin from the Truckee, Carson, Walker, and Humboldt Rivers resulted in an array of interconnected lake basins in relatively arid, low-lying valleys in western Nevada and eastern California (Morrison, 1991). Topographic enclosure of the lake basin, however, has obviated the development of a through-flowing river system. Limited basin interconnection linked to major flooding from lake overflow during late Cenozoic highstand conditions has been reported from several sites in the Great Basin, including: the Mono Basin in eastern California (Reheis et al., 2002); the Bonneville Basin in northern Utah (Gilbert, 1890; O'Connor, 1993), and the Alvord Basin in southeastern Oregon (Carter et al., 2006). The Owens (Jannik et al., 1991), Mojave (Meek, 1989; Miller, 2005), and Amargosa Rivers (Morrison, 1991; Menges and Anderson, 2005; Knott et al., this volume) provide examples of more extensive and persistent downstream-directed integration. In each of these cases, fluvial integration through once-enclosed basins continued to regions of insurmountable topographic enclosure, where the Amargosa River terminus in Death Valley is the best example. The Rio Grande and the Snake River provide two examples of complete downstream-directed integration to the sea (Mack et al., 1997; Connell et al., 2005; Wood and Clemens, 2002).

In each case cited here, the balance between water and sediment inputs and topographic impediments determined the extent, continuity, and persistence of integration. In the case of the Colorado River, the serial juxtaposition of large parts of the southern and central Rocky Mountains, the Colorado Plateau, the Basin and Range, and the Salton Trough appears to have provided an ultimately efficient conduit for connection of the upper basin with the sea. Perhaps the principal remaining questions involve the specific hydrologic and geologic circumstances on the Colorado Plateau and points upstream that combined to set the integration process in motion.

CONCLUSIONS

The new stratigraphic evidence and tephrochronologic data from Cottonwood and Mohave Valleys reported here are consistent with the lake-spillover model of Colorado River integration. The data, their implications, and their relations to other constraints on river evolution pose some challenges to models of river inception and early evolution that invoke combinations of subsidence, headward erosion, marine transgression, marine regression, and regional uplift. Our interpretation of the field evidence reported here is that a series of lakes developed after 5.6 Ma along the course of the lower Colorado River below the mouth of the Grand Canyon to at least Mohave Valley, and these lakes were drained in succession as divides were breached and lowered. Large volumes of coarse Colorado River sediment eventually arrived in these valleys, resulting in massive river aggradation between 5.6 and 4.1 Ma. The period of thick aggradation was concurrent with drainage integration and canyon enlargement upstream and with the arrival of Colorado River sediment in basins downstream. The proposed mode of basin interconnection and river integration through lake spillover is a well-documented phenomenon on numerous river systems in the interior of the western United States. The downstream integration of the lower Colorado River was particularly effective because it connected a large, well-watered highland source area with progressively lower-lying arid valleys and the developing Gulf of California.

APPENDIX 1. TEPHROCHRONOLOGY

The discovery and identification of three different late Cenozoic tephra beds at important stratigraphic levels in the study area provide new bracketing age controls on the evolution of the lower Colorado River. The tephrochronologic data help to establish a temporal context for our work that can be directly related to pre– and early Colorado River stratigraphic records reported from other sites both upstream and downstream of our study area.

Glass shards from eight samples of tephra layers in the study area were analyzed at the University of Utah Tephrochronology Laboratory. Analyses were done with a Cameca 50SX electron microprobe using methods discussed in Perkins et al. (1995, 1998). These analyses were compared with the laboratory's extensive database of electron microprobe analyses of late Cenozoic tephra layers in the western United States using methods of Perkins et al. (1998). Age estimates and errors for tephra lacking isotopic age control follow the methods discussed next. The age comparisons indicate that the tephra samples are each from one of three regionally distributed tephra layers: the ca. 3.3 Ma Nomlaki Tuff; the ca. 4.1 Ma lower Nomlaki tephra layer; or the 5.59 Ma tuff of Wolverine Creek (Table A1). We discuss chemical characteristics and age control for these three tephra next.

Tuff of Wolverine Creek

The tuff of Wolverine Creek was generated during one of a series of eruptions at ca. 5.6 to ca. 5.5 Ma in the Heise volcanic field in the eastern Snake River Plain (Morgan and McIntosh, 2005). The sequence of eruptions commenced with the emplacement of the tuff of Wolverine Creek and concluded with the emplacement of the Connant Creek Tuff. All these tuffs have essentially identical glass shard composition (Table A1) and are compositionally distinct from other silicic tuffs of the Heise volcanic field (Perkins and Nash, 2002). The single $^{40}Ar/^{39}Ar$ sanidine age for tuff of Wolverine Creek is 5.59 ± 0.05 Ma, while the weighted mean of three $^{40}Ar/^{39}Ar$ sanidine ages for the Connant Creek tuff and equivalent tuff of Elkhorn Spring is 5.51 ± 0.08 Ma (calculated from dates in Morgan and McIntosh, 2005).

Perkins et al. (1998) first identified a distal tuff from the tuff of Wolverine Creek–Connant Creek tuff sequence in the Muddy Creek Formation (Arrow Canyon section) of southern Nevada. Based on X-ray fluorescence (XRF) analysis, this distal tuff best matches analyses of the type tuff of Wolverine Creek. Compositionally similar tephra samples have been identified in the Muddy Creek Formation at Frenchman Mountain, Nevada (Castor et al., 2000; Castor and Faulds, 2001) and in the Lost Cabin beds in Cottonwood Valley along the lower Colorado River (House et al., 2005a). We conclude that it is likely that the Wolverine tephra is the only tephra from this Wolverine Creek–Connant Creek sequence in the Lower Colorado River region.

Lower Nomlaki Tephra Layer

The lower Nomlaki tephra layer is an informal name given to a tephra layer in Death Valley, California, which lies ~26 m below the Nomlaki tephra layer (Knott and Sarna-Wojcicki, 2001). The average glass shard composition of the lower Nomlaki tephra layer is similar to that of the Nomlaki tephra layer (discussed next), but it has measurably higher MnO and Cl and somewhat lower CaO values than the Nomlaki tephra layer. The lower Nomlaki tephra layer is further distinguished by a unimodal glass shard composition that contrasts with the polymodal glass shard composition of the Nomlaki Tuff (Table A1). Finally, the single mode of the lower Nomlaki tephra layer does not correspond to any of

TABLE A1. ELECTRON MICROPROBE ANALYSES OF LOWER COLORADO RIVER AND CORRELATIVE TEPHRA LAYERS

Sample/Tephra	Mod	*n*	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	MnO	MgO	CaO	BaO	K_2O	Na_2O	Cl	F	Sum	Du_1	Du_2	*D*avg	Comment
Wolverine Cr. avg.	–	20	74.06	.112	11.84	1.33	.033	.063	.495	.050	5.17	3.32	.128	.157	96.75	–	–	–	Weighted average
btr92-217	–	21	74.94	.112	11.84	1.30	.027	.050	.464	.046	5.06	3.46	.134	–	97.44	–	–	3.6	Ash-fall tuff between WOL & EHS?
acb92-6a	–	18	72.79	.117	11.65	1.30	.033	.066	.476	.033	5.46	2.91	.125	.122	95.09	2.9	–	1.7	Arrow Canyon location of Perkins et al. (1998)
btr92-215	–	22	74.64	.114	11.89	1.32	.039	.061	.481	.058	5.07	3.47	.131	–	97.27	1.0	–	1.1	Ash-fall tuff between WOL & EHS?
btr92-213	–	24	74.70	.110	11.90	1.35	.035	.062	.486	.062	5.12	3.43	.130	–	97.39	0.9	–	0.8	Type tuff of Wolverine Creek (WOL)
con94-677	–	19	70.79	.119	11.79	1.35	.035	.064	.490	.033	5.12	3.39	.124	.145	93.45	0.8	–	0.9	Type Connant Creek Tuff, basal ash fall
120304-6a	–	16	73.95	.123	11.64	1.35	.032	.065	.487	.046	5.12	3.30	.129	–	96.24	0.4	–	1.2	Odyssey Wash, Cottonwood Valley
btr92-218	I	19	74.55	.100	11.91	1.34	.036	.066	.505	.066	5.13	3.47	.133	–	97.30	2.0	–	1.4	Tuff of Elkhorn Spring? (EHS?); nonwelded
btr92-216	–	25	74.61	.117	11.89	1.32	.035	.064	.482	.060	5.09	3.47	.132	–	97.27	2.1	–	1.0	Ash-fall tuff between WOL & EHS?
btr92-214	–	21	74.55	.110	11.95	1.33	.039	.065	.486	.056	5.12	3.29	.131	–	97.14	0.6	–	0.8	Ash-fall tuff between WOL & EHS?
con94-679	–	10	76.27	.112	12.18	1.33	.031	.068	.505	.059	5.30	3.64	.122	.165	99.79	1.5	–	1.5	Type Connant Creek Tuff, ash-flow vitrophyre
jf99-455	–	20	73.55	.109	11.61	1.37	.034	.064	.514	.031	5.48	2.89	.131	.167	95.95	1.3	–	2.0	Frenchman Mountain (Castor et al., 2000)
jf-01-08	I	27	73.50	.110	11.76	1.28	.022	.059	.503	.043	5.29	3.00	.124	.179	95.85	2.7	–	1.6	Secret Pass Wash, Mohave Valley
btr92-219	I	21	74.49	.108	11.93	1.33	.029	.063	.546	.059	5.10	3.50	.122	–	97.28	3.5	–	4.1	Tuff of Elkhorn Spring? (EHS?); nonwelded
Nomlaki Mode IV	–	3	75.39	.132	11.47	0.78	.042	.121	.646	.092	4.06	3.29	.123	.054	96.20	–	29.5	–	Weighted average
PP041017-3	I	8	75.49	.138	11.44	0.78	.046	.119	.644	.086	4.08	3.30	.124	–	96.25	18.6	–	0.4	Silver Creek, Mohave Valley
Plush-Toy	IV	1	75.29	.127	11.58	0.83	.023	.131	.631	.075	4.46	2.86	.115	–	96.12	1.2	–	1.9	Silver Creek, Mohave Valley
pan93-332	III	1	74.69	.095	11.60	0.71	.031	.126	.681	.156	3.52	3.63	.121	.054	95.42	2.3	–	3.0	Panaca Fm., NV
Nomlaki Mode I	–	12	75.10	.195	12.01	0.97	.039	.162	.860	.074	3.59	3.62	.121	.052	96.79	–	18.2	–	Weighted average
pan93-332	II	4	74.58	.183	11.97	0.92	.031	.146	.829	.071	4.04	3.26	.122	.110	96.26	5.8	–	3.2	Panaca Fm., NV
wlw02-1236	I	11	74.67	.195	11.86	0.97	.059	.156	.826	.074	3.36	3.71	.116	.067	96.07	1.5	–	2.1	Willow Wash, CA
PP041017-3	II	7	74.87	.189	11.75	0.98	.042	.177	.847	.081	3.65	3.63	.117	–	96.34	2.1	–	2.0	Silver Creek, Mohave Valley
Plush-Toy	II	7	75.45	.187	11.98	1.03	.033	.166	.842	.062	3.49	3.46	.113	–	96.82	1.6	–	2.0	Silver Creek, Mohave Valley
nt91-01	I	21	74.66	.199	12.02	0.99	.035	.161	.876	.063	3.58	3.64	.125	.065	96.41	2.2	–	1.0	Type Nomlaki
nt91-02	I	19	75.90	.199	12.19	0.95	.035	.164	.880	.088	3.64	3.68	.124	.017	97.87	1.2	–	1.4	Type Nomlaki
Lower Nomlaki	–	19	73.16	.212	12.75	1.09	.072	.219	.909	.102	3.30	3.74	.166	.075	95.79	–	15.8	–	Weighted average
topock#1	–	19	73.25	.203	12.65	1.08	.068	.215	.838	.055	3.50	3.75	.161	.073	95.85	7.1	–	3.9	Powerline, Mohave Valley
CWAJB02-1	–	15	73.34	.202	12.76	1.07	.072	.211	.914	.116	3.50	4.02	.167	.115	96.49	4.1	–	1.2	Wishy wash, Cottonwood Valley, NV
jf-01-07	–	19	72.90	.206	12.70	1.06	.064	.215	.904	.105	3.03	3.83	.171	.056	95.24	0.7	–	0.9	13th Green tephra bed, Bullhead City, AZ
CAES#1-1143	–	22	73.33	.222	12.80	1.09	.075	.222	.923	.127	3.04	3.44	.165	.068	95.50	1.7	–	1.0	Core sample, Bristol Lake, CA
JRK-DV-39	–	22	73.02	.223	12.84	1.12	.076	.229	.956	.104	3.47	3.74	.167	.074	96.03	2.2	–	3.0	Type Lower Nomlaki
Nomlaki Mode II	–	11	73.56	.230	12.45	1.19	.046	.224	1.121	.066	3.46	3.61	.115	.081	96.15	–	22.9	–	Weighted average
PP041017-3	III	6	73.86	.225	12.14	1.14	.042	.209	1.064	.088	3.50	3.63	.123	–	96.03	5.4	–	3.4	Silver Creek, Mohave Valley
wlw02-1236	II	6	73.77	.235	12.27	1.16	.054	.220	1.066	.072	3.21	3.82	.122	.085	96.08	0.9	–	2.8	Willow Wash, CA
nt91-01	II	8	73.88	.224	12.50	1.15	.055	.221	1.111	.062	3.45	3.52	.114	.078	96.35	1.9	–	1.2	Type Nomlaki
nt91-02	II	10	74.85	.240	12.56	1.16	.048	.221	1.106	.074	3.49	3.81	.116	.032	97.71	0.8	–	1.1	Type Nomlaki
Plush-Toy	I	8	74.44	.214	12.46	1.23	.042	.233	1.108	.066	3.35	3.56	.113	–	96.82	2.1	–	1.8	Silver Creek, Mohave Valley
pan93-332	I	28	72.65	.233	12.49	1.21	.044	.227	1.156	.057	3.52	3.54	.112	.099	95.34	2.4	–	1.8	Panaca Fm.
Nomlaki Mode III	III	2	73.07	.277	12.80	1.41	.038	.277	1.290	.071	3.21	3.63	.113	.062	96.25	–	15.6	–	Weighted average
wlw02-1236	III	2	72.98	.317	12.63	1.41	.050	.264	1.244	.050	3.11	3.84	.112	.086	96.09	5.4	–	2.7	Willow Wash, CA
pan93-332	III	2	71.42	.298	12.55	1.44	.025	.268	1.278	.079	3.01	3.75	.116	.067	94.32	1.1	–	1.5	Panaca Fm.
nt91-01	III	2	73.04	.245	12.97	1.36	.030	.277	1.306	.044	3.28	3.00	.129	.050	95.73	2.0	–	1.9	Type Nomlaki
Plush-Toy	III	4	73.72	.248	12.85	1.43	.038	.286	1.292	.091	3.25	3.74	.106	–	97.05	1.5	–	1.4	Silver Creek
nt91-02	III	1	74.05	.332	13.11	1.36	.053	.293	1.360	.070	3.46	3.81	.110	.030	98.04	2.5	–	4.0	Type Nomlaki

Notes: Mode—for samples with more than one compositional mode, modes are labeled "I", "II", etc., from most abundant to least abundant mode, while samples with only a single mode are not labeled unless the mode shows a compositional range, in which case the mode is labeled "R". *n*—number of shards analyzed for sample or mode. Total Fe as Fe_2O_3. Du_1, Du_2, *D*avg—statistical distances using TiO_2, Fe_2O_3, MgO, CaO, and Cl. Du_1 is from a particular sample to the next overlying sample (excluding tephra averages); Du_2 is the distance from one tephra average to the next overlying tephra average; *D*avg is the distance from each tephra sample to the tephra average. Ideally, as discussed in Perkins et al. (1995, 1998), these distance statistics are expected to have a chi-square distribution for 5 degrees of freedom. Thus, in general, 95% of statistically identical analyses will lie within 3.3 distance units of one another. However, the model used to estimate analytical precision is based on average precision, and individual analyses can deviate from model values used in calculations in this table. All analyses are given in weight percent. Analyses of lower Colorado River tephra are highlighted with italic font; sample weighted averages are shown in bold. Cr.—Creek; EHS—Elk Horn Spring; Fm.—Formation; WOL—Wolverine Creek.

the modes recognized in the Nomlaki Tuff, so microprobe analyses can be used to confidently distinguish between the lower Nomlaki and Nomlaki tephra layers.

Our age estimate for the lower Nomlaki tephra layer is 4.06 ± 0.46 (1σ) Ma, or, rounding to significant decimal places, 4.1 ± 0.5 Ma. This age is the weighted mean of two linear extrapolation age estimates using Bristol Lake core CAES#1 (Sarna-Wojcicki et al., 2001) and Artist Drive section 2 (Knott and Sarna-Wojcicki, 2001). Error estimates for individual extrapolation ages are from an empirical error model for such age estimates (M. Perkins, 2007, personal commun.). This model, fitted to tephra-bearing sections and cores throughout the Basin and Range, conservatively reflects observed variation in sedimentation rates in these sections and cores.

The lower Nomlaki tephra layer has a reversed polarity (Knott and Sarna-Wojcicki, 2001). Thus, it likely lies within either chron C2Ar (3.58–4.18 Ma), as suggested by Knott and Sarna-Wojcicki (2001), or within chron C3n.1r (4.29–4.48 Ma). An improved estimate of the age of this tephra layer is needed in order to distinguish between these two possibilities.

The mean glass shard compositions of individual Lower Nomlaki tephra samples vary more than expected based on analytical error alone. It is uncertain if this variation represents variability within the lower Nomlaki tephra layer or if it indicates that there are, perhaps, two or more separate tephra layers with similar but somewhat different compositions. Tephra samples jf-01–07 and CWAJB02 from the lower Colorado Basin are compositionally identical to CAES#1–1143 in the Bristol Lake boring CAES#1. In contrast, sample JRK-DV-39 from the type area of the lower Nomlaki tephra layer, has measurably lower CaO values than these three samples. Similarly, Topock#1 from the lower Colorado Basin also is measurably different from these first three samples and has measurably lower CaO values. For the present, we conclude that all five samples are most likely from a single tephra layer, but future findings may require modification of this conclusion.

Nomlaki Tuff

The 3.3 Ma Nomlaki Tuff is a widespread tephra layer (Sarna-Wojcicki et al., 1991). With a source in the Lassen Peak area of northeastern California, the Nomlaki Tuff is recorded in many areas of California, including Death Valley. It is also present in the Bonneville Basin of Utah (Williams, 1994), east Central Nevada (M. Perkins, 2007, personal commun.), and as far southeast as the Rio Grande Rift (Connell et al., 1999). As reported by Williams (1994) and Knott and Sarna-Wojcicki (2001), the Nomlaki Tuff was deposited during the Mammoth subchron of the Gauss polarity chron. Williams (1994), based on sedimentation rates at the base of the Burmester core, estimated an age of 3.29 Ma for the Nomlaki Tuff. Our age estimate for the Nomlaki tephra layer is 3.29 ± 0.05 Ma using the error model discussed previously. This is a weighted average of two extrapolation age estimates using the Burmester, Utah, core (Williams, 1994) and the Willow Wash, California, section (Reheis et al., 1991).

Glass shards of the Nomlaki tephra layer are compositionally distinctive and readily distinguished from those of other tephra layers in the database. As first recognized by Williams (1994), glass shards of the Nomlaki tephra generally fall within one of three compositional modes: mode I, the low Fe_2O_3 mode (~0.98 wt%); mode II, the intermediate Fe_2O_3 mode (~1.18 wt%); or mode III, the high Fe_2O_3 (~1.40 wt%) mode (Table A1). All Nomlaki Tuff samples have two abundant modes, modes I and II. Mode III is missing in some analyses, but this likely reflects the low frequency of this mode. Furthermore, no other tephra layers are know to contain these three modes. Thus, modes I and II and mode III (when present) are characteristic of the Nomlaki Tuff. Finally, we note that a fourth, very low Fe_2O_3 (~0.75 wt%) mode, mode IV, is observed in samples from the lower Colorado River region and east central Nevada (Table A1). Since these samples also have the characteristic modes (I, II, and III) of the Nomlaki Tuff, we are confident they are samples of the Nomlaki Tuff.

ACKNOWLEDGMENTS

Our research has benefited greatly from interactions with many individuals, including Jon Spencer, Jim Faulds, John Bell, Keith Howard, Lee Amoroso, Sue Beard, Tracey Felger, Daniel Malmon, Amy Brock, Brenda Buck, Joel Pederson, Karl Karlstrom, Jon Patchett, Andy Cohen, Dave Kimbrough, and Simon Hook. We thank Andrei Sarna-Wojcicki, U.S. Geological Survey, for providing samples of the lower Nomlaki tephra layer from the Artist Drive area and from the Bristol (dry) Lake core CAES#1. We also thank Barbara Nash, University of Utah, for providing the analyses of glass shards for Nomlaki Tuff samples collected in the type area by Andrei Sarna-Wojcicki. Our ongoing research in the Lower Colorado River area has been supported largely by various grants from the USGS STATEMAP program to the Nevada Bureau of Mines and Geology and Arizona Geological Survey. The National Park Service provided occasional logistical support. This manuscript was improved by constructive reviews from Richard Hereford, Daniel Malmon, Jeffrey Knott, and David Miller.

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Manuscript Accepted by the Society 17 July 2007

Printed in the USA