



Lacustrine cave carbonates: Novel archives of paleohydrologic change in the Bonneville Basin (Utah, USA)

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ABSTRACT

Records of past changes in lake levels and lake water isotopic composition in closed basins provide key insights into past variations in the hydrological cycle; however, these records are often limited by dating precision and temporal resolution. Here we present data from lacustrine cave carbonates, a previously unexplored class of carbonates that comprise a promising new archive of past hydrologic changes in the Bonneville Basin of the northeastern Great Basin (USA). These dense carbonates precipitated within caves, crevices, and other protected spaces flooded by Lake Bonneville during its highstand in the last glacial period. We report on deposits in Cathedral and Craners caves, located ~50 km apart at similar elevations approximately 100 m above the modern Great Salt Lake and almost 200 m below Lake Bonneville's highstand shoreline. Carbonates from the two caves show similar chronologies, mineralogical transitions, isotopic compositions, and uranium concentrations. These findings suggest that lacustrine cave carbonates record changes in lake level and in the isotopic composition and chemistry of lake water. Importantly, the deposits can be precisely dated by U–Th methods, providing the first records of Lake Bonneville's water balance changes tied to precise U–Th ages. Close agreement between paired U–Th and calibrated ¹⁴C ages in the deposits suggests a minimal (< 200 a) carbon reservoir effect in the lake and allows ¹⁴C dating to be used for age control in portions of the deposits less suitable for U–Th dating.

We use ages for the onset and cessation of lacustrine cave carbonate deposition to offer new constraints on past changes in lake level and the carbonate saturation state of lake water. We also present precisely dated, high-resolution oxygen and uranium isotope records from the deposits. Within a first phase of deposition reflecting the lake's transgression between 26 and 18 ka, our isotopic data suggest a large influx of freshwater during Heinrich Stadial 2. A hiatus in deposition beginning at 18.2 ± 0.3 ka may be the result of freshening related to the lake's overflow. Calcite deposition resumes at Cathedral Cave at 16.4 ± 0.2 ka, suggesting that basin overflow had ceased by this time and that the lake's calcite saturation state had increased, and $\delta^{18}\text{O}$ values increase markedly after 15.9 ka, consistent with drying at this time. These data imply that the lake's deglacial regression began well before the Bølling warming. Cessation of this second phase of deposition at 14.7 ± 0.2 ka, coincident with the Bølling warming, may reflect the lake's drop below Cathedral Cave's elevation. A final stage of aragonite deposition between 13.8 and 13.6 ka differs from the underlying facies and may not reflect deposition from lake waters filling the cave.

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1. Introduction

More than a century of study of lacustrine deposits in the Great Basin (western USA) has established that the last glacial period was marked by dramatic expansions of terminal lakes (e.g., Gilbert, 1890; Antevs, 1948; Eardley et al., 1957; Broecker and Kaufman, 1965; Benson et al., 1990; Oviatt et al., 1992).

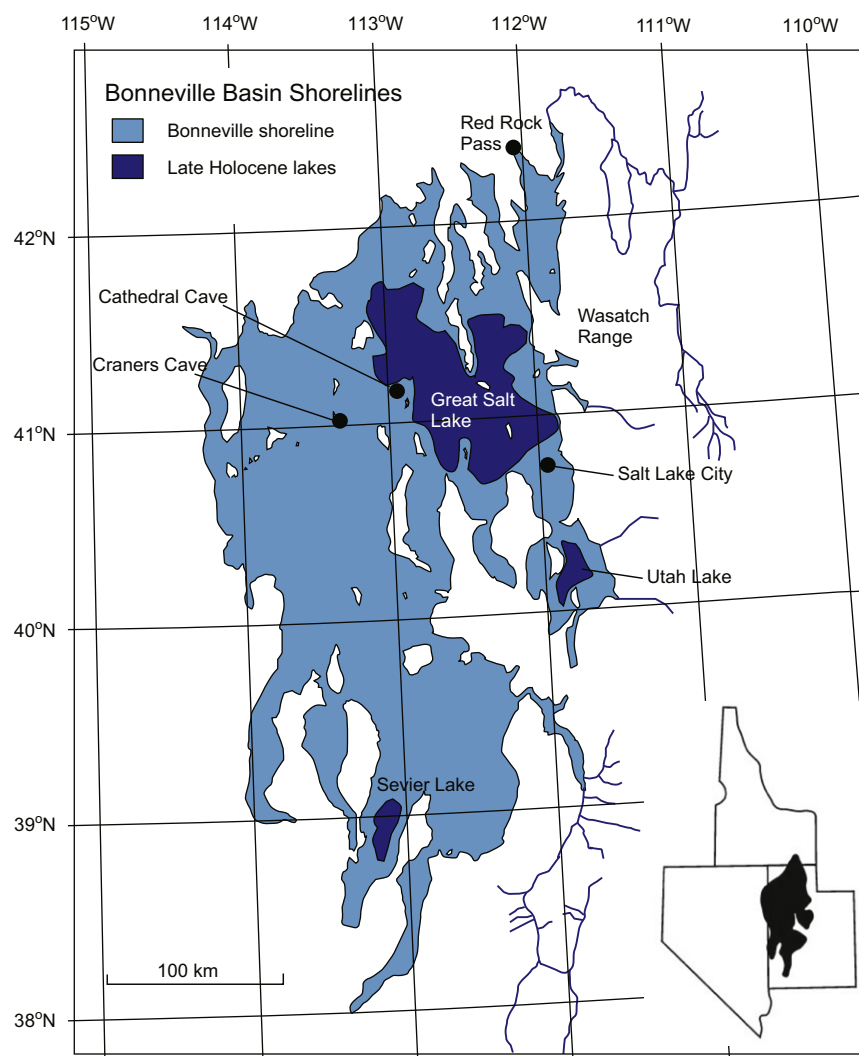


Fig. 1. Regional map showing sample locations, major shorelines and key geographic features in the Bonneville Basin. Red Rock Pass is the site of overflow at the Bonneville and Provo levels. Modified from Quade and Broecker (2009) and Nelson et al. (2005). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

The largest of these lakes, Lake Bonneville in the northeastern Great Basin, spanned more than 50,000 km² at its maximum extent, an area approximately a factor of ten greater than that of the modern Great Salt Lake and other Late Holocene lakes in the Bonneville Basin (Fig. 1).

Records of past lake-level changes provide key insights into changes in the balance of precipitation and evaporation (P–E) in the western US, and in particular into the past behavior of the winter storm track, the dominant control on the region's water balance (Cayan and Peterson, 1989; Lall and Mann, 1995; Mann et al., 1995; Moon et al., 2008). Lake-level records offer a potentially valuable target for climate models exploring the last glacial period and deglaciation and could allow reconstruction of shifts in atmospheric circulation associated with abrupt climate changes such as Heinrich events and the Bølling warming. The relatively low resolution and dating precision of most records, however, make it difficult to determine the timing and rates of water balance changes and their relationships to abrupt climate changes recorded elsewhere. Currently, well-resolved records from the basin disagree as to the sign of hydrologic change during Heinrich events. Oviatt (1997) suggested that ice rafting events coincide with increased P–E in the basin, while Benson et al. (2011) concluded that Heinrich events are marked by drier

conditions. These uncertainties limit our understanding of the region's water balance history, our ability to reconstruct atmospheric circulation changes associated with abrupt climate changes, and the utility of Lake Bonneville records for modeling efforts.

Here we present high-resolution, precisely dated records from a novel paleohydrologic archive: lacustrine cave carbonates. These deposits were first recognized during excavations in Cathedral Cave near the center of the Bonneville Basin (Madsen et al., 2000), where they form a coating up to 10 cm thick on the cave walls, ceiling, and floor. Our work represents the first in-depth study of these deposits. We find that these carbonates, found throughout the basin in caves and other protected spaces formerly flooded by the lake, record changes in lake level, the oxygen and uranium isotope composition of lake water, and lake water chemistry. Importantly, the age of the deposits can be precisely determined by U–Th methods, presenting a unique opportunity to obtain lake records anchored to precise ages.

This paper focuses on introducing this new class of carbonates and establishing them as a promising new archive of lake conditions. We also explore implications from two cave records for the timing of Lake Bonneville's rise and fall during the last

glacial period and termination. These deposits appear to be common in the Bonneville Basin and hold considerable potential as paleoclimatic archives.

2. Setting, materials and methods

2.1. Lake-level changes in the Bonneville Basin

Beginning after 30 ka, Lake Bonneville underwent a prolonged transgression from low levels (likely < 1300 m above sea level, similar to the elevation of the modern Great Salt Lake) that continued through the last glacial maximum (LGM; 19–23 ka) (Fig. 2) (Oviatt et al., 1992). The transgression was punctuated by 3–5 small (tens of m) regressions (Currey, 1990; Oviatt et al., 1990; Oviatt et al., 1992; Oviatt, 1997; Sack, 1999; Patrickson et al., 2010). During the first and largest regression, the Stansbury oscillation, the Stansbury shoreline near 1350 m was formed at ~25 ka (Oviatt et al., 1990). After 18.6 ± 0.4 ka (15.25 ± 0.16 ^{14}C ka) (Oviatt, 1991), Lake Bonneville reached its highstand and began to overflow at least intermittently into the Snake River basin, stabilizing the lake at the Bonneville shoreline (1552 m) (Currey, 1982). The alluvial fan damming the lake failed sometime before 17.4 ± 0.4 ka (14.3 ^{14}C ka) (Oviatt et al., 1992), causing the lake to catastrophically flood the Snake River basin (Malde, 1968; Jarrett and Malde, 1987). Recently obtained ages of gastropods in beach deposits post-dating the flood suggest a minimum age of 18.1 ka for the flood (Miller et al., in press). The lake then stabilized at the new threshold elevation and carved the prominent Provo shoreline (1444 m) (Currey, 1982).

It is thought that the lake overflowed at least intermittently at the Provo level for 1–3 ka before beginning its deglacial regression, but the timing of the regression – and its relationship to the Bølling warming at 14.7 ka – is highly uncertain (Oviatt et al., 1992; Godsey et al., 2011). Ages on offshore deposits suggest that the regression began at 15 ± 0.5 ka (Godsey et al., 2011), while

ages within Provo beach deposits extend only to ~17 cal ka, suggesting an earlier regression (Miller et al., in press).

The lake regressed to below 1290 m sometime before 13.0 ka. During oscillations between 12.9 and 11.2 ka, approximately the time of the Younger Dryas cold period, the lake rose to ~1295 m and formed the Gilbert shoreline (Oviatt et al., 2005). The lake has been lower than 1287 m throughout the Holocene Epoch and maintained an average level of 1283 m during the past century (Murchison, 1989; Oviatt et al., 1992).

2.2. Lacustrine cave deposits

We have found that many caves and protected spaces in the central and western Bonneville Basin below the highstand shoreline are coated with dense, layered carbonates with thicknesses of 1–10 cm (Fig. 3). We have sampled these deposits in over forty caves, crevices, and interstitial spaces in gravel deposits at elevations ranging from 15 m above the modern lake to 50 m below the Bonneville shoreline. This paper focuses on deposits in Cathedral Cave in the Lakeside Range and Craners Cave in the Newfoundland Range (Fig. 1; Table 1).

These lacustrine cave deposits constitute a new class of terrestrial carbonates clearly distinguishable from tufas and speleothems. Lake Bonneville tufas are porous and dull in appearance and grow in branching, cauliflower-like forms. Lacustrine cave deposits are dense, with almost no pore spaces, have a vitreous appearance and tend to form continuous sheets of carbonate coating cave and crevice walls (Fig. 3). The deposits range in texture from fine (~0.1 mm) horizontal layers to radiating fibrous crystals to overlapping mammillary growths. Their morphology and patterns of occurrence distinguish them from vadose-zone speleothems and flowstones.

Samples were collected from Cathedral and Craners Caves in August 2008. Both caves are located within the Utah Test and Training Range (UTTR) to the west of the Great Salt Lake. Cathedral Cave is a solution cavern formed in the Permian Oquirrh Limestone in the Lakeside Range at a modern elevation

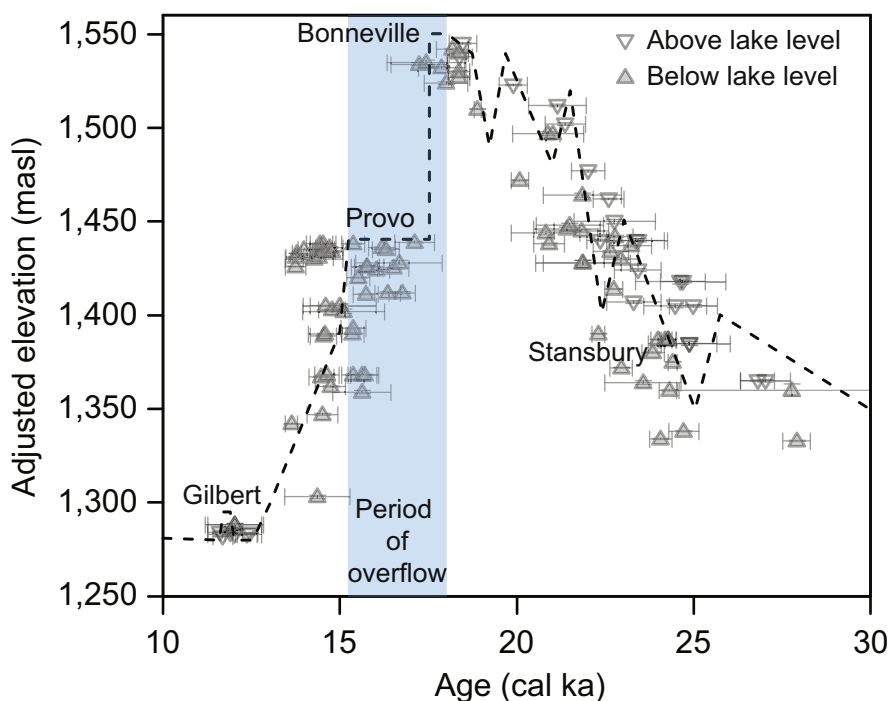


Fig. 2. Schematic lake level history for Lake Bonneville based on previous work. Also shown are calibrated radiocarbon ages limiting lake elevation ($\pm 1\sigma$) (grey triangles) and period of open hydrologic conditions (blue shading) (Oviatt et al., 1992; Godsey et al., 2005, 2011; C.G. Oviatt, pers. comm.).

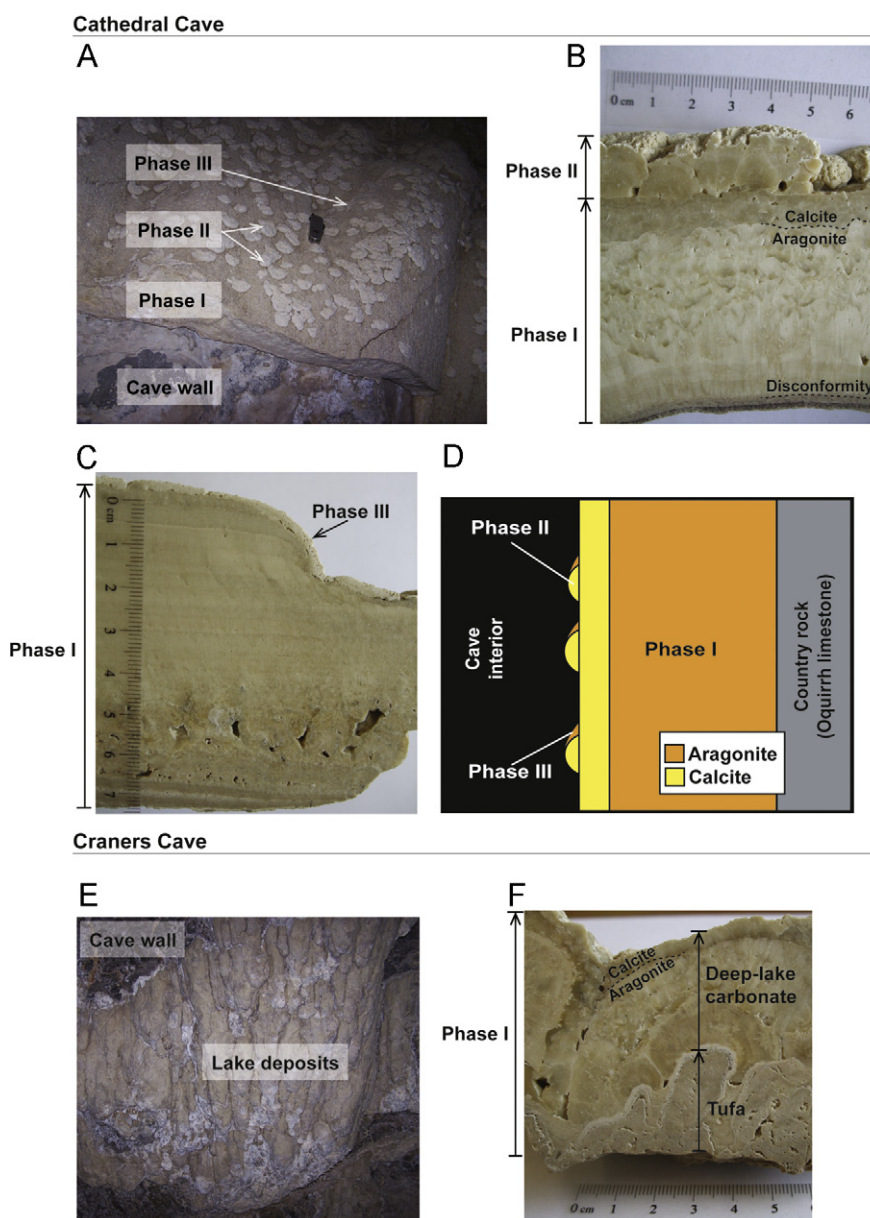


Fig. 3. Cathedral Cave (A–D) and Craners Cave (E–F) deposits. (A) Cathedral Cave Phase I–III deposits on cave wall. (B) Polished section showing Phases I and II deposits with disconformity and transition from aragonite to calcite within Phase I. (C) Cathedral Cave Phases I and III deposits. (D) Schematic stratigraphy of Cathedral Cave deposits. (E) Craners Cave deposits on cave wall. Note the undulating surface due to underlying photic zone tufa. Area shown is approximately 3 m across. (F) Craners Cave Phase I deposits showing opaque, irregular photic zone tufa at base overlain by cave carbonate indicative of deep-water deposition. The transition from aragonite to calcite is also shown. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Table 1
Cave elevations.

Name	Elevation (masl)	Corrected elevation (masl)	Latitude (°N)	Longitude (°W)
Cathedral Cave	1393	1359	41.1653	112.9358
Craners Cave	1386	1358	41.0191	113.3519

of 1393 m. Tufas associated with the Stansbury shoreline are present at the entrance. The cave appears to predate the Bonneville lake cycle (Madsen et al., 2000). The cave walls are nearly vertical and are coated with carbonate (Fig. 3A). Coatings near the front of the cave include branching tufa overlain by dense carbonate, while coatings at the back consist entirely of dense carbonate. Samples for this study were collected from cave walls

~60–65 m from the cave entrance at an elevation of approximately 1383 m (Madsen et al., 2000).

Craners Cave (informally named) lies in the Ordovician Fish Haven Dolomite in the Newfoundland Mountains ~50 km southwest of Cathedral Cave. The cave is situated within the Stansbury shoreline complex at a modern elevation of 1386 m. The cave contains a main room ~30 m wide, 20 m deep, and up to 10 m high with four small (~3 m wide × 3 m deep) rooms extending into the back wall. Cave walls are nearly vertical, and the cave ceiling is composed of cemented beach gravel. Our interpretation is that the cemented gravel pile predates the last lake cycle and initially covered an overhang in the country rock. The cave was then excavated by wave action during the early portion of the last transgression. Branching heads of photic zone tufa line the walls and are overlain by dense cave carbonate (Fig. 3E,F). Samples were taken from the back wall of

the cave, ~20 m from the entrance, at an approximate elevation of 1382 m.

Both Cathedral and Craners Caves are very dry. There are no speleothems or other evidence of dripping water except immediately at the cave entrances. The floors of both caves are covered with packrat and raptor middens, which are not found in wet caves. We are therefore confident that the sampled lacustrine cave carbonates have not been exposed to liquid water since the regression of the paleolake.

2.3. Isostatic rebound corrections

Elevations were determined at cave entrances by Trimble GeoXH handheld GPS. Isostatic rebound of the basin following the drying of Lake Bonneville left the basin's shorelines up to 75 m higher in the basin's center than at its edges (Currey, 1982). To compensate for isostatic effects, adjusted elevations are calculated using a site's modern vertical distance from the lowest point in the basin and the modern elevation of the local highstand (Bonneville) shoreline (Oviatt et al., 1992). Using this approach, the current sampling elevation of Cathedral Cave at 1383 m is estimated to have been 1351 m during the last lake cycle, within the established elevation range of the Stansbury shoreline complex. For Craners Cave, the present sampling elevation of 1386 m is estimated as 1358 m prior to rebound.

Bills and May (1987) suggest that isostatic compensation in the Bonneville Basin occurs mostly over the first 1 ka after the lake has attained a new level. When the lake began rising, the caves would have been near their modern elevation, and the lag associated with isostatic compensation would have delayed the lake's rise through the caves' elevation. When the lake began falling from the Provo shoreline, the caves would have been ~30–50 m below their adjusted elevations, delaying the lake's fall past the caves.

2.4. Sampling and analysis

Samples were slabbed, polished, and washed with ultrapure water and ethyl alcohol prior to sampling. Powders for stable isotope analysis were drilled at 0.5 mm intervals in Cathedral Cave samples (~75 a resolution) and 2 mm intervals (~200–300 a resolution) in Craners Cave samples. Carbon and oxygen isotopic compositions of samples were measured using an automated carbonate preparation device (KIEL-III) coupled to a gas-ratio mass spectrometer (Finnigan MAT 252) at the University of Arizona. Powdered samples were reacted with dehydrated phosphoric acid under vacuum at 70 °C. The isotope ratio measurement is calibrated based on repeated measurements of NBS-19 and NBS-18. Precision is $\pm 0.11\%$ for $\delta^{18}\text{O}$ and $\pm 0.08\%$ for $\delta^{13}\text{C}$ (1 sigma). All oxygen and carbon isotopic values are expressed in per mil (‰) in delta notation relative to the VPDB standard.

The dominant carbonate mineralogy of Cathedral Cave samples was determined using a Philips X'pert X-ray Diffractometer (PW3040-MPD) located at Rutgers University. Analyses followed the methods of Chiu et al. (2005).

Strontium isotope ratios ($^{87}\text{Sr}/^{86}\text{Sr}$) were measured in three samples of Phase I carbonate from Cathedral Cave. Powdered samples were dissolved in a ~1.4 M solution of ultrapure HNO_3 in Teflon beakers. The samples were dried on a hotplate and redissolved in 3.5 M HNO_3 . The solution was then run through Eichrom Sr-specific resin columns to isolate strontium. Samples were analyzed for $^{87}\text{Sr}/^{86}\text{Sr}$ on a thermal ionization mass spectrometer (TIMS) at the University of Arizona Geochronology Laboratory. NBS-987 standards were analyzed on a daily basis for normalization, yielding an $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.710258 ± 0.000021 ($n = 13$).

Samples analyzed for Mg concentrations were removed from the CC3 hand sample using a microdrill with a carbide steel bit.

Aliquots of ~2–3 mg were drilled in a profile perpendicular to growth bands, immediately adjacent to stable isotope and U/Th sample sites. Sample dissolution and standard preparation procedures were performed in a class-100 clean lab using Teflon prepared by multistep boiling-washing in concentrated aqua regia, HCl, and HNO_3 . Mg was calibrated using multi-element standards of 1.25, 10, 50, 100, and 500 ppb solutions. Ca was calibrated using 0.1, 0.5, 1, 10, 50, 100, and 200 ppm standards prepared by dissolving reagent grade calcite powder. Concentrations were measured using a Thermo Scientific Element2 single-collector ICP-MS, using ^{24}Mg and ^{44}Ca in low-resolution mode. We confirmed that interferences from doubly charged ^{48}Ca were negligible by measuring ^{24}Mg in medium resolution mode. Mg concentrations in CaCO_3 were calculated by assuming stoichiometric proportionality between Ca and CaCO_3 in each sample.

Samples for ^{14}C dating were extracted and cubed with a diamond saw, then washed with distilled water. The whole, unground cube was used for dating. All pretreatments were performed at the University of Arizona, followed by conversion of samples to graphite. Measurements were conducted by the University of Arizona Accelerator Mass Spectrometer Facility. For conversion of ^{14}C ages (expressed as ^{14}C ka) to calendar ages (expressed as cal ka), we used the IntCal04 calibration curve (Reimer et al., 2004) with OxCal 4.1 (Bronk Ramsey, 2009) due to the closer correspondence of IntCal04 to more recent estimates of atmospheric ^{14}C evolution between 17.5 and 14.5 ka (Southon et al., 2012). For two samples with calibrated ages extending beyond 26 ka, we used the IntCal09 calibration curve (Reimer et al., 2009).

U–Th dating was performed at the University of Minnesota. Sample powders were drilled using a carbide-tipped drill bit. Sample sizes were 5–20 mg in aragonite portions of the deposits and 20–100 mg in calcite portions. Samples were prepared following the methods of Cheng et al. (2000, 2009) and Shen et al. (2002) and analyzed using a Neptune multi-collector ICP-MS or Element single-collector ICP-MS in peak-jumping mode. Uncertainties on concentrations are estimated as $\pm 1\%$ to reflect uncertainties in spike concentration and weighing. Chemical blanks were measured with each set of 10–15 samples and were

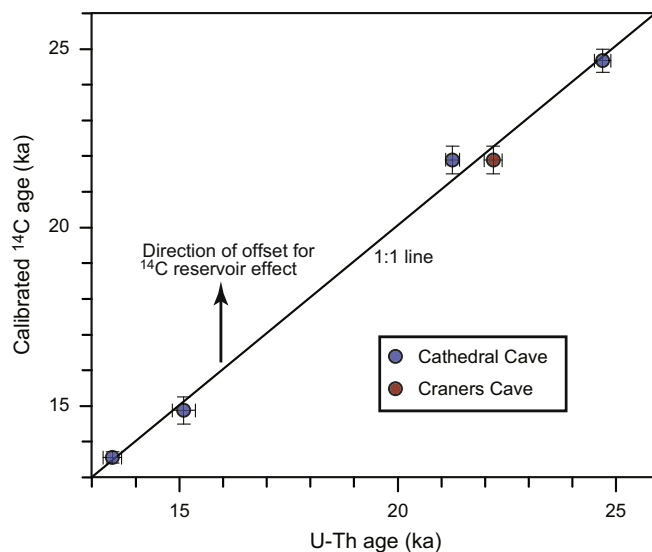


Fig. 4. U–Th ages and calibrated ^{14}C ages from paired samples from Cathedral and Craners Caves. A 1:1 line is shown for comparison. The data suggest a minimal reservoir effect for dissolved inorganic carbon in the lake between 25 and 13 ka. Error bars reflect 2σ confidence intervals. Calibration of ^{14}C ages uses IntCal04.

negligible (< 0.05 fg ^{230}Th ; < 0.8 fg ^{234}U ; < 0.5 pg ^{232}Th and ^{238}U) for all samples. Ages were calculated using the ^{230}Th and ^{234}U half-lives of Cheng et al. (submitted) and the ^{238}U half-life of Jaffey et al. (1971). The lower-precision ^{230}Th and ^{234}U half-lives in Cheng et al. (2000) produce ages within our stated uncertainties. A detrital $^{230}\text{Th}/^{232}\text{Th}$ atomic ratio of $(8 \pm 2) \times 10^{-6}$ was used for samples from both caves to estimate contributions of initial ^{230}Th . The method used to determine this ratio is described in the Supplementary text.

3. Results

3.1. U–Th and ^{14}C ages

We obtained 51 ^{234}U – ^{230}Th ages of carbonates from Cathedral and Craners Caves, which range from 13.6 ± 0.1 to 25.5 ± 0.4 ka (Supplementary Table 1). The deposits are primarily composed of aragonite with high U concentrations (8–22 $\mu\text{g/g}$) and high $^{230}\text{Th}/^{232}\text{Th}$ ratios. As a result, aragonite samples commonly have 2σ uncertainties of ± 60 – 100 a. Calcite portions of the deposits have lower U concentrations (< 2 $\mu\text{g/g}$), causing corrections for initial ^{230}Th to be relatively larger and reducing the precision of U–Th ages in calcite samples.

We obtained 19 ^{14}C ages from these deposits, ranging in age from 11.7 ^{14}C ka to 21.3 ^{14}C ka (Supplementary Table 2). The aridity of the caves and the dense, pristine appearance of the cave carbonate suggests that this material does not suffer from the contamination problems of common lake tufas (Benson and Thompson, 1987; Benson, 1993).

We also obtained paired U–Th and ^{14}C ages from 7 samples (Supplementary Table 2). In comparing the values we exclude 2 pairs from Phase I calcite, as these values were used in determining the detrital $^{230}\text{Th}/^{232}\text{Th}$ ratio (Supplementary Text). In 4 of the 5 remaining pairs, calibrated ^{14}C ages agree within 2σ uncertainties with U–Th ages (Fig. 4). This agreement suggests that the lake's ^{14}C reservoir effect was less than 200 a, as previously suggested by Oviatt et al. (1992). Because of this minimal ^{14}C reservoir effect, we use calibrated ^{14}C ages for age control in portions of the deposits with high detrital ^{230}Th contamination (base and top of Phase I). Age models for $\delta^{18}\text{O}$ records were constructed using linear interpolation (Supplementary Text; Supplementary Table 3).

Deposits at Cathedral Cave show three depositional phases described in detail below. At Craners Cave we have identified deposits that appear similar in morphology and stratigraphic position to the first two phases at Cathedral Cave, but to this point we have only dated and studied samples from Phase I.

3.2. Phase I

The first phase of deposition in Cathedral Cave is composed of 4–7 cm of aragonite overlain by 0.5–1 cm of calcite. The bottom 3–7 mm of the aragonite consists of a set of mm-scale horizontal laminae with an upper surface that is irregular and easy to separate from the remainder of the deposit, suggesting that it is bounded by a disconformity. Above this likely disconformity, the deposit varies between fine (sub-mm) laminae and intergrown mammillary forms. Sparry crystals are present in a few samples (Fig. 3C) and were avoided. The transition from aragonite to calcite is evident in a distinct change of color and luster (Fig. 3B).

U–Th ages in the aragonite portion of Phase I progress in stratigraphic order, beginning with a near-basal age of 25.5 ± 0.4 ka. A ^{14}C age on basal deposits gives a similar result: 26.0 ± 0.8 ka (21.72 ± 0.22 ^{14}C ka). A U–Th age just above the hypothesized disconformity indicates that it ended before 24.89 ± 0.08 ka, after which aragonite deposition continued until 18.83 ± 0.05 ka. Following the transition to calcite, deposition continued until 18.3 ± 0.3 ka (14.98 ± 0.08 ^{14}C ka).

At Craners Cave, the basal deposits consist of 2–5 cm of photic-zone tufa with a branching form and dull luster. Translucent, vitreous aragonite with a thickness of ~ 5 cm overlies the tufa. The aragonite occurs both as mm-scale laminae and cm-scale radiating fibrous crystals. Calcite approximately 1 cm thick overlies the aragonite and is distinguished by a change in color (Fig. 3F).

A ^{14}C age from the Craners Cave basal tufa provides a minimum age of 26.0 ± 0.7 ka (21.67 ± 0.19 ^{14}C ka) for the lake's rise past this elevation. The U–Th age closest to the base of the dense aragonite gives a minimum age of 23.5 ± 0.2 ka for the beginning of deep-water carbonate precipitation. A U–Th age at the top of the aragonite deposits indicates that the transition to calcite deposition occurred just after 18.88 ± 0.08 ka, while ages from the outer edge of the calcite indicates that deposition stopped at 18.1 ± 0.3 ka (U–Th); a ^{14}C age from the outer edge calibrates to 18.1 ± 0.4 ka (14.86 ± 0.04 ^{14}C ka).

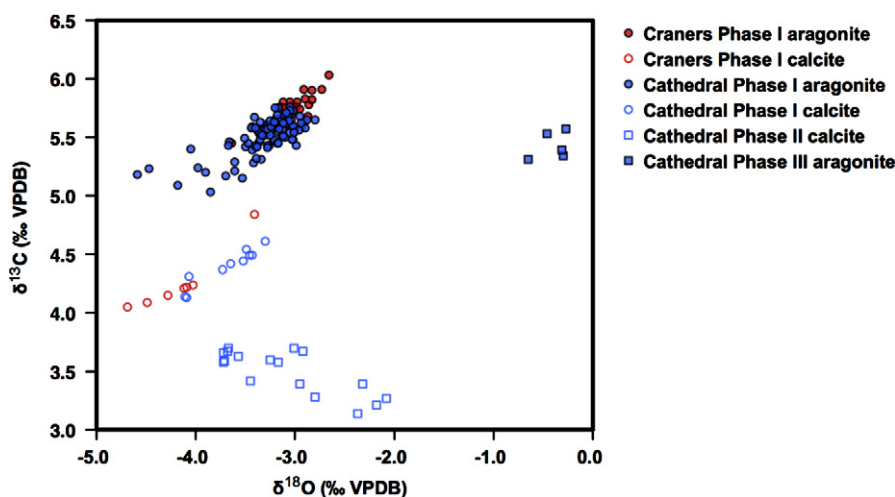


Fig. 5. Stable isotope data from all phases of deposition at Cathedral and Craners Caves. $\delta^{13}\text{C}$ data clearly distinguish between calcite and aragonite. $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values are positively correlated in Phases I and III, but are negatively correlated in Phase II. $\delta^{18}\text{O}$ values are anomalously positive in Phase III.

At Cathedral Cave, we observe that deposits at the front of the cave consist of tufa overlain by dense, evenly layered deposits, while deposits at the back of the cave are entirely dense and layered. Based on this observation we suggest that the basal tufa forms in the presence of light, perhaps due to association with microbial or algal communities. As Craners Cave is not as deep as Cathedral Cave, and light could have reached the back of the cave, we suggest that basal tufa at Craners Cave began growing when the lake was in shallow water. This tufa was subsequently covered by dense carbonate when the cave was in deep water after 23.5 ka.

$\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data from Phase I in the two caves form overlapping arrays that demonstrate positive correlations between O and C isotope ratios (Fig. 5). Values in Craners Cave vary over a larger range than in Cathedral Cave. Calcite is clearly distinguished from aragonite by its less positive $\delta^{13}\text{C}$ values.

3.3. Phase II

Phase II consists of discrete mammillary calcite growths up to 2 cm in diameter. Nesting dome-shaped depositional horizons are visible in cut sections. This phase has only been studied in detail in samples from Cathedral Cave, though morphologically and stratigraphically similar features have been identified at Craners Cave.

U–Th ages from Phase II deposits indicate that deposition occurred between 16.4 ± 0.2 and 14.7 ± 0.2 ka. $\delta^{18}\text{O}$ data extend to less negative values than in Phase I, while $\delta^{13}\text{C}$ values are substantially less positive than Phase I (Fig. 5). In contrast to the strong positive correlations in Phase I, $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values are negatively correlated in this phase.

3.4. Phase III

Phase III consists of a discontinuous layer of porous, dull aragonite that is up to 8 mm thick at the front of Cathedral Cave and 2–4 mm thick near the back of the cave. Unlike the other two phases, Phase III drapes over horizontal surfaces and does not appear on vertical or overhanging faces. U–Th ages indicate that it formed between 13.8 ± 0.1 and 13.6 ± 0.1 ka. $\delta^{18}\text{O}$ values are 2–4‰ higher than in other phases, while $\delta^{13}\text{C}$ values are similar to Phase I aragonite (Fig. 5).

4. Discussion

4.1. Lacustrine cave carbonates as archives of lake water isotopic composition and chemistry

During the last large lake cycle in the Bonneville Basin, abundant carbonates formed as shorezone tufas and as fine-grained endogenic carbonate that precipitated in surface waters and accumulated in sediments (Oviatt et al., 1994; Nelson et al., 2005; Benson et al., 2011). The widespread formation of these carbonates indicates that the lake was commonly saturated with respect to carbonate species, a conclusion supported by modeling of lake chemistry (Spencer et al., 1985).

We find that the pure, dense carbonates described in this paper are confined to caves and other protected spaces. We suggest that caves, crevices and interstitial spaces in gravel deposits provided the nucleation sites, low detrital inputs and still-water conditions necessary for dense carbonate precipitation from saturated lake waters. The lacustrine carbonate only formed when two conditions were met: when paleolake waters inundated the cave, and when that lake water was sufficiently

saturated with respect to calcium carbonate to precipitate calcite or aragonite.

Our data suggest that lacustrine cave carbonates reflect the isotopic and chemical composition of lake waters and are negligibly affected by local factors such as groundwater or interaction with bedrock. Although Cathedral and Craners caves are ~50 km from each other and lie within different country rock, the change from aragonite to calcite and the cessation of the main phase of deposition occur synchronously within the uncertainty of our age models at both caves. Additionally, U concentrations and variations in O and U isotopes are extremely similar in both caves' deposits (Fig. 6). $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data from the caves also closely match the range of values found in shoreline tufas (Nelson et al., 2005) and are more positive than would be expected if groundwater derived. As described in Section 3.1, paired U–Th and ^{14}C ages from both caves indicate a minimal reservoir age for carbon in the deposits, suggesting negligible local inputs from limestone country rock (Fig. 4). Finally, $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in Cathedral Cave Phase I deposits range from 0.7116 to 0.7118. These values are consistent with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios inferred for Lake Bonneville's waters from tufas and shells (0.7114–0.7118) (Hart et al., 2004) and are distinct from the likely composition of the Paleozoic limestone country rock (< 0.709).

4.2. Reconstructing water balance from lacustrine cave carbonates

4.2.1. Top and bottom ages

If it is accepted that these deposits formed when lake water was both above a given cave's elevation and sufficiently saturated with respect to calcite or aragonite, bottom ages of deposits may record either the lake's transgression through a cave's elevation or an increase in the carbonate saturation state of lake waters. Top ages may record the opposite: either a regression below the cave's elevation or a decline in the carbonate saturation state of lake waters due to freshening.

4.2.2. Oxygen isotope ratios

The deposits provide precisely dated, high-resolution $\delta^{18}\text{O}$ records that reflect changes in the isotopic composition of lake water. The $\delta^{18}\text{O}$ of lake waters is largely determined by the basin's water balance (Talbot, 1990), as precipitation adds low- ^{18}O waters and evaporation enriches lake waters in ^{18}O . Modern precipitation in the basin has a weight-averaged $\delta^{18}\text{O}$ of ~−15‰ (VSMOW) (Bowen et al., 2007), and fossil mammal bones from Wyoming suggest that regional precipitation $\delta^{18}\text{O}$ was at least as negative during the last glacial period (Kohn and McKay, 2010). Bonneville tufas and lacustrine cave carbonates have $\delta^{18}\text{O}$ values of −3‰ to −5‰ (VPDB) (Nelson et al., 2005; this study). Even assuming lake temperatures of 0 °C (a possibility for deep lake waters precipitating lacustrine cave carbonates), these values require that lake water $\delta^{18}\text{O}$ values were ~7‰ less negative than precipitation $\delta^{18}\text{O}$ due to evaporative enrichment of ^{18}O . Variations in this enrichment due to changes in P–E would have had substantial leverage to change lake water $\delta^{18}\text{O}$.

Lake water $\delta^{18}\text{O}$ may also be impacted by changes in the relative contributions of different drainages and in the isotopic composition of precipitation. Sr isotope ratios measured in lake carbonates indicate that the relative importance of drainages did not radically change between ~26 and ~14 ka (Hart et al., 2004).

The influence of changes in the isotopic composition of precipitation on lake water $\delta^{18}\text{O}$ can be estimated using regional speleothem records. Calcites in Devils Hole (southern Nevada) suggest a ~1‰ increase in $\delta^{18}\text{O}$ between 30 and 10 ka (Winograd et al., 2006). A speleothem record from Arizona indicates a decrease of 1–2‰ in precipitation $\delta^{18}\text{O}$ during Heinrich events

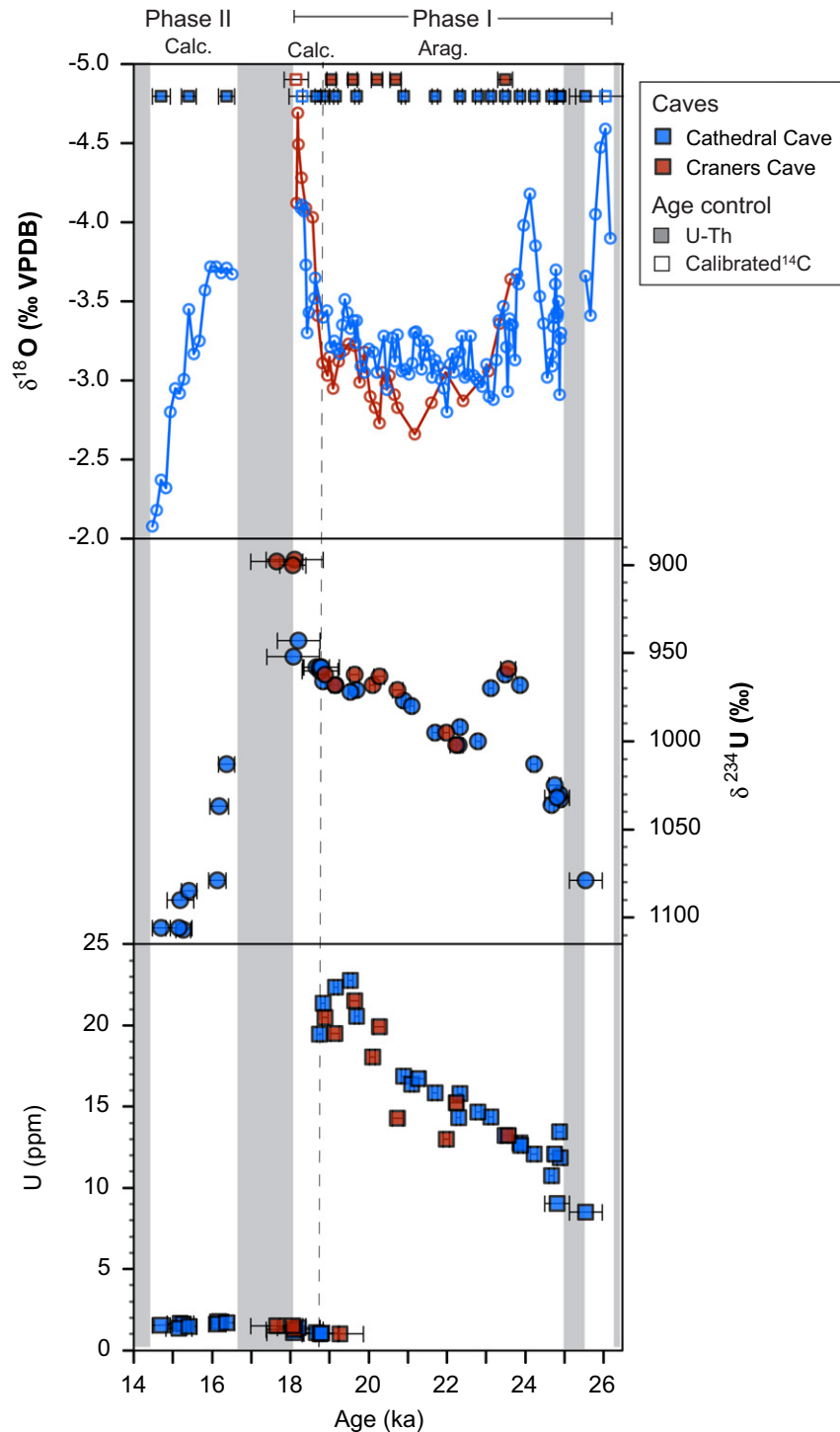


Fig. 6. Comparison of data from Cathedral (blue) and Craners (red) Caves. Depositional periods and dominant mineralogy are indicated at the top, and periods of non-deposition are shaded. *Top:* Oxygen isotope data from the two caves, with age control points and their 2σ uncertainties shown above; shaded boxes indicate U–Th ages, and open boxes indicate calibrated ^{14}C ages. *Middle:* $\delta^{234}\text{U}$ data from the two caves. *Bottom:* U concentration data. O and U isotopes and U concentrations show only minor differences between the two caves. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

and an increase of 2‰ at the Bølling warming (14.7 ka) (Wagner et al., 2010). Similarly small amplitudes of change are observed in a speleothem from the western edge of the Bonneville Basin recording the penultimate deglaciation (Shakun et al., 2011) and in a speleothem from southern Nevada that spans the beginning of Heinrich 1 (Lachniet et al., 2011). These small changes suggest that precipitation $\delta^{18}\text{O}$ has less influence on lake water $\delta^{18}\text{O}$ than do changes in P–E.

To estimate residence times for Lake Bonneville's waters, we use lake volumes estimated for the lake at the Stansbury and Bonneville shorelines (Nelson et al., 2005) and modern input fluxes estimated by Arnow (1980). Based on the assumption that stream and spring inputs were no more than a factor of 5 higher during the transgression and highstand, we estimate water residence times of 100–800 yr. The century-scale residence time of lake waters smooths the impact of water balance changes on lake water $\delta^{18}\text{O}$ values,

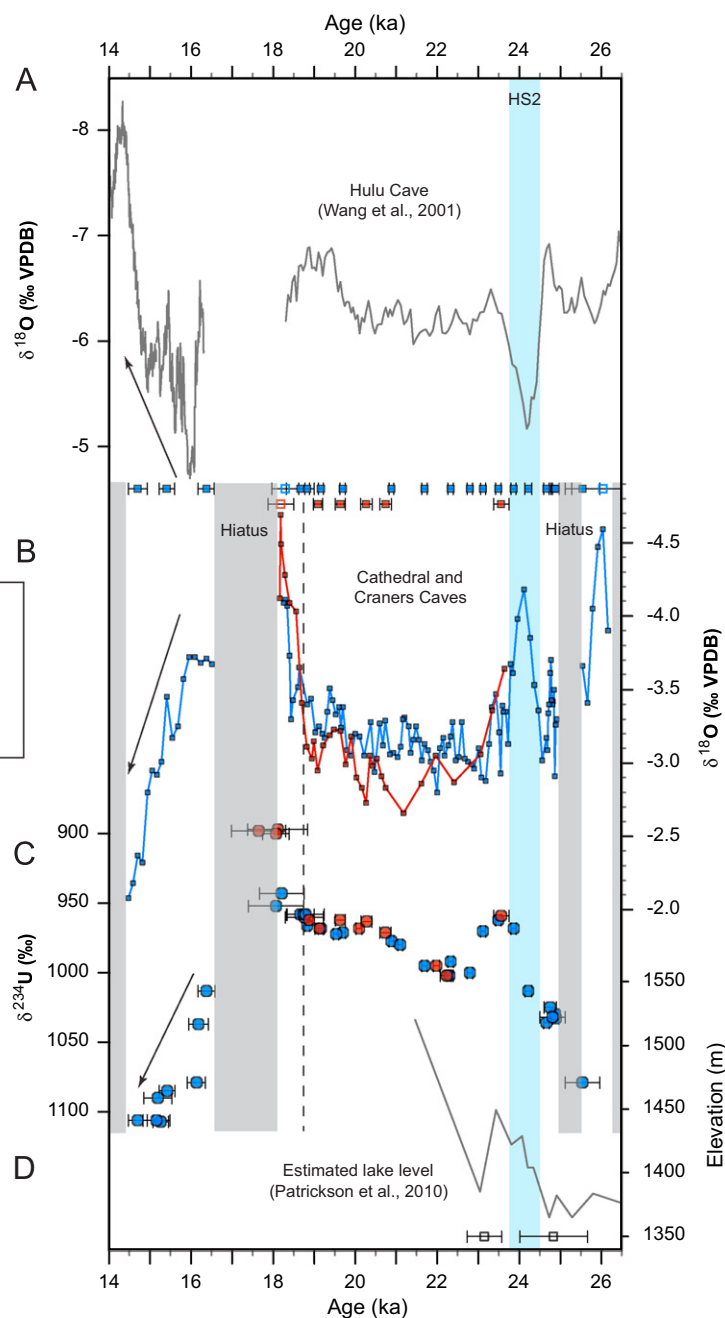


Fig. 7. Cathedral and Craners Cave records of Lake Bonneville's isotopic composition from 26 to 14 ka. (A) $\delta^{18}\text{O}$ record from Hulu Cave stalagmites H82 and MSD shown as an independent U–Th dated estimate both of the age of Heinrich Stadial 2 (HS2; blue bar) and of a gradual warming in the Northern Hemisphere beginning at 16 ka (Wang et al., 2001). (B) $\delta^{18}\text{O}$ records from Cathedral and Craners Cave deposits (blue and red, respectively). Age control points with 2σ uncertainties are shown above the records; shaded boxes indicate U–Th ages, and unshaded boxes indicate calibrated ^{14}C ages. Also shown are periods of no deposition (grey boxes) and the aragonite–calcite transition in Phase I (dashed line). (C) $\delta^{234}\text{U}$ data from Cathedral and Craners Cave with 2σ age uncertainties. (D) Lake Bonneville lake level reconstruction of Patrickson et al. (2010) with ^{14}C age control points below, showing lake level rise during HS2 at the same time as negative excursions in $\delta^{18}\text{O}$ and $\delta^{234}\text{U}$ in Cathedral Cave deposits. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

making the direction and rate of change in $\delta^{18}\text{O}$ values better indicators of water balance than the absolute values.

4.2.3. $^{234}\text{U}/^{238}\text{U}$ ratios in lacustrine cave carbonates

Complementary hydrologic information may be provided by normalized $^{234}\text{U}/^{238}\text{U}$ activity ratios ($\delta^{234}\text{U}$) in these deposits. $\delta^{234}\text{U}$ of dissolved lacustrine U may plausibly be linked to two related variables: the balance between groundwater and surface waters reaching the lake, and the intensity of chemical weathering.

^{238}U is typically held in stable sites within minerals, while ^{234}U is more susceptible to weathering due to being located in radiation-damaged sites and to being recoiled out of the grain. As a result, river and groundwaters tend to have elevated $\delta^{234}\text{U}$ due to preferential leaching of ^{234}U (Chabaux et al., 2003). Deeply penetrating groundwaters tend to have especially high $\delta^{234}\text{U}$ due to interaction with unweathered rock with abundant leachable ^{234}U ; $\delta^{234}\text{U}$ tends to be lower in shallow-sourced waters due to their interaction with weathered soils that have already lost ^{234}U (Riotte and Chabaux, 1999). As a result, $\delta^{234}\text{U}$ of lacustrine U may reflect the balance of shallow- and deep-sourced inputs. $\delta^{234}\text{U}$

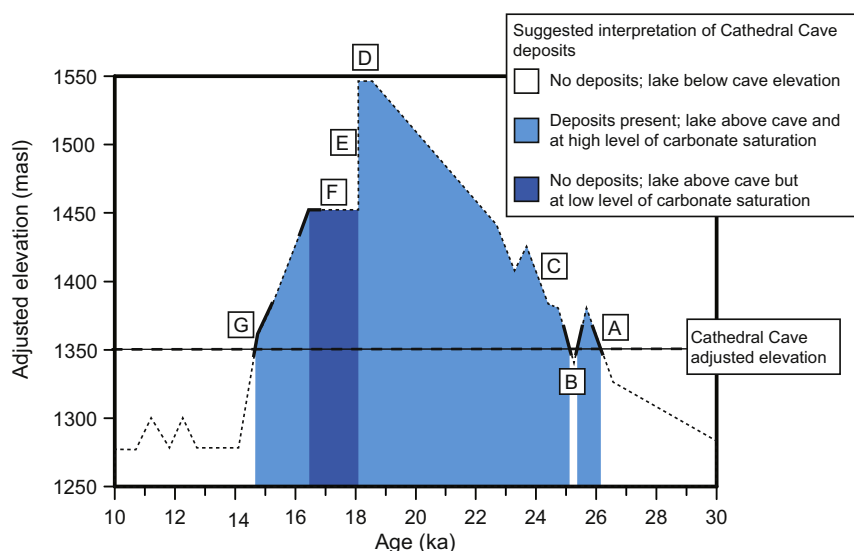


Fig. 8. Schematic hydrograph of lake levels in the Bonneville basin based on our interpretation of Cathedral Cave deposits. Periods of deposition are here assumed to reflect times when the lake was both higher than the cave's elevation and at high levels of calcite or aragonite saturation. Solid portions of the curve are more confidently constrained by lacustrine cave carbonates, while dotted portion are more uncertain. (A) Initial rise of the lake past Cathedral Cave's elevation; (B) hiatus during first Stansbury Oscillation; (C) fast lake rise during Heinrich Stadial 2 indicated by $\delta^{18}\text{O}$ and $\delta^{234}\text{U}$ data; (D) onset of overflow at the Bonneville shoreline; (E) Bonneville flood; (F) overflow at the Provo shoreline; (G) deposition of Phase II during the lake's fall from the Provo shoreline past the cave's elevation. In the diagram, the end of Phase I deposition coincides with the Bonneville flood, but this relationship is uncertain.

may also respond to changes in chemical weathering, as increased chemical weathering more congruently removes ^{234}U and ^{238}U , lowering $\delta^{234}\text{U}$. Accordingly, $\delta^{234}\text{U}$ values of dissolved riverine U are inversely related to precipitation rates in drainages on New Zealand's South Island (Robinson et al., 2004), and changes in speleothem $\delta^{234}\text{U}$ have been interpreted as reflecting chemical weathering changes (e.g., Hellstrom and McCulloch, 2000).

During periods of increased P–E in the Bonneville Basin, the proportion of deep-sourced groundwater inputs may decrease as runoff increases. Chemical weathering may also increase with increasing P–E (though temperature changes should also play a role). Both changes would cause $\delta^{234}\text{U}$ values of U entering the lake to decrease with rising P–E. Consistent with this expectation, $\delta^{234}\text{U}$ values decrease during the lake's rise between 25 and 18 ka and rise during the lake's fall after 16.4 ka (Figs. 6 and 7).

Dissolved U in Lake Bonneville waters is likely to have a much longer residence time than the waters themselves, delaying and smoothing $\delta^{234}\text{U}$ changes with respect to $\delta^{18}\text{O}$ changes.

4.3. New insights into Lake Bonneville's transgression: 26–18.2 ka

Lacustrine cave deposits at Cathedral Cave began forming by 26.0 ± 0.8 ka (21.72 ± 0.22 ^{14}C ka), and tufa began forming in Craners Cave at the same time (26.0 ± 0.7 ka; 21.67 ± 0.19 ^{14}C ka), indicating that the lake had risen above the caves' elevations near the Stansbury shoreline by this time (Fig. 8A).

A possible disconformity follows in Cathedral Cave deposits, consistent with sedimentological data indicating a lake regression (the "Stansbury oscillation") at this time (Fig. 8B) (Currey et al., 1983; Oviatt et al., 1990; Patrickson et al., 2010). We were unable to test for a synchronous hiatus in Craners Cave deposits due to the irregular stratigraphy of the tufas that comprise this interval (Fig. 3F). Deposition resumes by 24.89 ± 0.08 ka, marking the end of the oscillation; this timing is consistent with an independent age for shoreline deposits from the Stansbury oscillation (24.7 ± 0.8 ka; 20.7 ± 0.3 ^{14}C ka) (Currey et al., 1983).

$\delta^{18}\text{O}$ values decrease by $\sim 1\%$ from 24.5–24.0 ka, suggesting an increase in P–E. The $\delta^{18}\text{O}$ decrease correlates in time with a

rise in lake level inferred from lake sediments (Patrickson et al., 2010) (Figs. 7D and 8C). A decrease in $\delta^{234}\text{U}$ values also begins at 24.5 ka (Fig. 7C), consistent with an increase in shallow-sourced inputs or increased chemical weathering accompanying increased precipitation. This $\delta^{234}\text{U}$ decrease lags and lasts longer than the $\delta^{18}\text{O}$ decrease, reflecting the longer turnover time for dissolved U in lake waters.

The decrease in $\delta^{18}\text{O}$ and $\delta^{234}\text{U}$ starting at 24.5 ka corresponds in time with the age of Heinrich Stadial 2 (HS2) indicated by the Hulu Cave record (Fig. 7) (Wang et al., 2001). $\delta^{18}\text{O}$ values in our record then begin rising again at 24.0 ka, in agreement with the end of HS2 in Hulu Cave. This correlation suggests that HS2 was marked by an increase in P–E in the basin, perhaps due to a strengthening of the Northern Hemisphere subtropical jet and a corresponding southerly displacement of the storm track in association with the southward displacement of the ITCZ. In effect, this response would be a mirror image of the Southern Hemisphere storm track response to Heinrich events suggested by hosing experiments (Lee et al., 2011).

$\delta^{18}\text{O}$ values then increase after 24.0 ka, reaching a maximum at approximately 23.2 ka (Fig. 7). This period of more negative P–E may correspond to a regression evident in the stratigraphic record (Fig. 7D) with an age of 23.0 ± 0.4 ka (19.37 ± 0.08 ^{14}C ka; Patrickson et al., 2010). $\delta^{234}\text{U}$ values also become greater but again lag the $\delta^{18}\text{O}$ changes.

$\delta^{18}\text{O}$ values decrease by $\sim 1.2\%$ after ~ 19 ka, at approximately the same time as the transition to calcite (18.8 ka) (Figs. 6 and 7). Approximately half this change ($\sim 0.6\%$) may reflect the change in the fractionation of ^{18}O associated with the transition from aragonite to calcite (Kim and O'Neil, 1997; Kim et al., 2007). However, while $\delta^{13}\text{C}$ values and U concentrations change immediately at the transition, $\delta^{18}\text{O}$ values do not (Figs. 6 and 7). This observation suggests that the $\delta^{18}\text{O}$ shift at least partly reflects a change in the isotopic composition of lake water, likely due to more positive P–E after 19 ka.

$\delta^{234}\text{U}$ values decrease after 21.5 ka, stay approximately constant from 20.5 to 19 ka, then again decrease (Fig. 7). The magnitude of the decrease after 19 ka is quite different in the

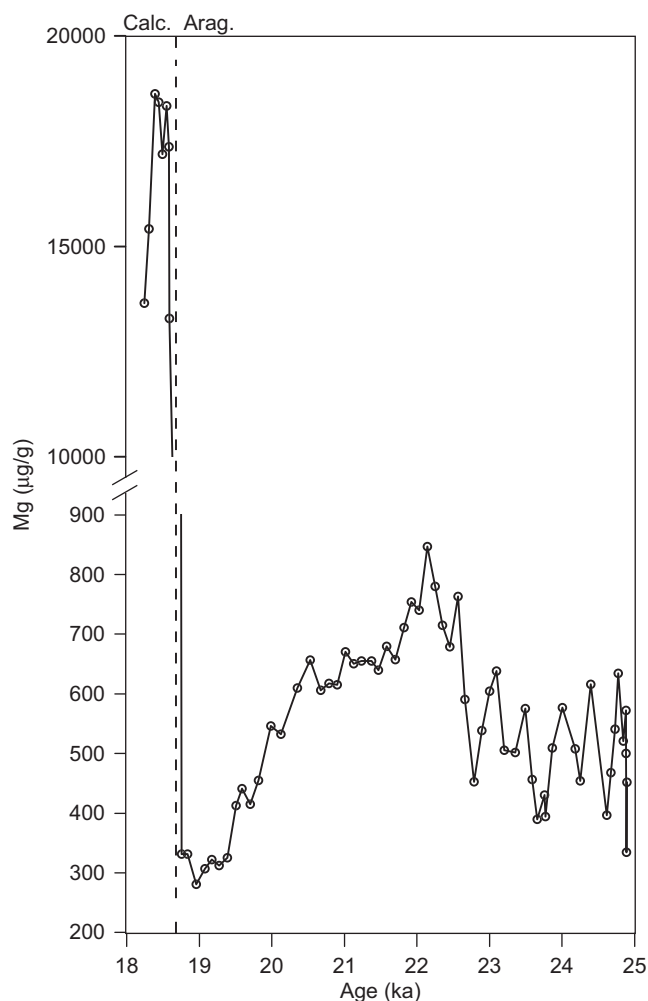


Fig. 9. Mg concentrations in Phase I carbonate from Cathedral Cave. Mg concentrations drop to their lowest levels just before the transition from aragonite to calcite, suggesting that the mineralogical change may reflect declining Mg/Ca ratios in lake water. The abrupt rise in Mg concentrations after the transition reflects the higher distribution coefficient of Mg in calcite. Note the axis break and change of scale in the vertical axis.

two caves, with $\delta^{234}\text{U}$ values falling to 900‰ in Craners Cave but only to 940‰ in Cathedral cave.

We suggest that the decrease in $\delta^{18}\text{O}$ values at 19.0–18.8 ka (Fig. 7) may mark the onset of overflow at the Bonneville shoreline (Fig. 8D), which would have decreased water residence time in the lake. This interpretation is consistent with the best constraining age for the shoreline, a ^{14}C age on charcoal in a pre-Bonneville soil just below the shoreline of 18.6 ± 0.4 ka (15.25 ± 0.16 ^{14}C ka) (Oviatt, 1991).

The transition from aragonite to calcite after 18.8 ± 0.1 ka in the Cathedral and Craners Cave deposits is consistent with the reduction in aragonite/calcite ratios in endogenic carbonate in lake sediments prior to the lake's highstand (Spencer et al., 1984; Oviatt et al., 1994; Benson et al., 2011). This mineralogical transition may be driven by declining Mg/Ca ratios during the lake's rise. Mg/Ca ratios are thought to be an important determinant of carbonate mineralogy, with high ratios favoring aragonite deposition (Hardie, 1996; Morse et al., 1997; Davis et al., 2000). Modeling of the lake's chemistry indicates that Ca concentrations were buffered by calcium carbonate precipitation, while Mg concentrations more nearly tracked salinity, leading to falling Mg/Ca ratios during the lake's transgression (Spencer et al., 1985).

To test whether Mg/Ca ratios in lake waters may have contributed to this mineralogical change, we measured Mg concentrations in Cathedral Cave Phase I deposits (Fig. 9). Mg levels range between ~400–600 ppm between 25 and 22.5 ka, then rise to ~800 ppm by 22 ka. This rise may reflect dissolution of playa evaporites as the lake expanded or higher Mg concentrations in the lake's deep waters as the cave was submerged in the rising lake. After 22 ka, Mg levels fall, reaching a minimum just before the transition to calcite. This finding is consistent with a relationship between lake water Mg/Ca and the mineralogy of these deposits. After the transition to calcite, Mg levels in the deposit rise to >10,000 ppm due to the higher distribution coefficient of Mg in calcite.

Deposition in both Cathedral and Craners Caves ceases at 18.2 ± 0.3 ka (14.98 ± 0.08 ^{14}C ka and 14.86 ± 0.04 ^{14}C ka, respectively). This cessation may reflect the reduction in salinity and calcite saturation state resulting from overflow. Dating of the Provo shoreline suggests that the catastrophic drop from the Bonneville shoreline (the "Bonneville flood"; Fig. 8E) has a minimum age of 18.1 ka (Miller et al., in press). This correspondence suggests that the end of Phase I deposition could be related to the flood. As the caves were ~200 m beneath the lake surface at the Bonneville stage, it is possible that the cessation of deposition reflects freshening of deep lake waters in the caves by mixing associated with the flood.

4.4. New insights into Lake Bonneville's regression: 16.4–14.7 ka

Phase II, a set of dome-shaped layers composed entirely of calcite, formed between 16.4 ± 0.2 and 14.7 ± 0.2 ka. We hypothesize that deposition of Phase II started after the lake ceased overflowing at the Provo shoreline, began concentrating solutes and attained higher calcite saturation (Fig. 8F). Phase II ended as the lake fell below the cave's elevation (Fig. 8G). $\delta^{18}\text{O}$ values are roughly constant in the first 0.5 ka of Phase II before beginning to rise at 15.9 ka, suggesting that pronounced drying began at this time (Fig. 7). $\delta^{234}\text{U}$ also rises during this phase (Fig. 7), consistent with reduced chemical weathering accompanying drying and/or an increase in groundwater inputs as the lake fell.

If our interpretation is correct, our data suggest that drying of the basin began well before the Bølling warming (14.7 ka) at roughly the same time that well-dated climate archives in China, the Cariaco Basin, and Greenland begin indicating the first sustained deglacial warming of the Northern Hemisphere (~16 ka) (Grootes et al., 1993; Grootes and Stuiver, 1997; Peterson et al., 2000; Wang et al., 2001). Our interpretation splits the "Mystery Interval" (17.5–14.5 ka) (Denton et al., 2006) in the Bonneville Basin into two portions: a wetter first portion (beginning after ~19 ka) characterized by fresh, at least intermittently overflowing conditions, and a drier second portion (beginning at 16.4 ± 0.2 ka). This is opposite to the pattern at Lake Estancia in New Mexico of a "Big Dry" between ~18.2 and 16.0 cal ka followed by a highstand from ~16.0–14.7 ka (Allen and Anderson, 2000), which was suggested by Broecker et al. (2009) as a template for the hydrologic imprint of the Mystery Interval in the western U.S. However, a speleothem record from Fort Stanton Cave does not indicate a similar dry-wet pattern, despite its proximity to Lake Estancia (Asmerom et al., 2010); the period from 18.2–16.4 ka has generally low $\delta^{18}\text{O}$ values, interpreted as wet conditions, followed by higher $\delta^{18}\text{O}$ from 16.4–15.4 ka. Meanwhile, wet conditions are indicated throughout the Mystery Interval by a speleothem from Cave of the Bells, Arizona (Wagner et al., 2010). Directly west of the Bonneville Basin, Lake Lahontan appears to have begun falling just before 15.5 ± 0.3 ka (13.1 ^{14}C ka) (Adams and Wesnousky, 1998). The Cathedral Cave

record suggests that Lake Bonneville's regression began almost 1000 years earlier, though the beginning of the $\delta^{18}\text{O}$ rise at 15.9 ka is closer to the timing of the Lahontan record. Clearly, more work is needed to understand spatial and temporal patterns in hydrological changes in the western U.S. during the last glacial period and deglaciation.

A final period of deposition (Phase III) occurred between 13.80 ± 0.12 and 13.57 ± 0.13 ka. Unlike the first two phases, Phase III deposits are only found on subhorizontal surfaces at Cathedral Cave, are thicker at the front of the cave than at the back, are porous and dull, and have not been conclusively identified at other locations. Additionally, the deposits have $\delta^{18}\text{O}$ values $\sim 2\text{--}4\text{‰}$ less negative than other phases. These observations suggest a different origin from Phases I and II.

5. Conclusions

Lacustrine cave carbonates in the Bonneville Basin provide records of lake level and lake water isotopic composition and chemistry tied to precise U–Th and ^{14}C ages. These records lay the groundwork for new investigations of controls on the basin's water balance and the timing of hydrologic changes in the western U.S. during the last glacial period and deglaciation. Data from Cathedral Cave suggest that Heinrich Stadial 2 was marked by an increase in P–E in the Bonneville Basin. Cathedral Cave deposits also suggest that Lake Bonneville began drying by 16.4 ± 0.2 ka, with $\delta^{18}\text{O}$ values rising substantially beginning at 15.9 ka, well before the Bølling–Allerød warm period.

Lacustrine cave carbonates have been identified in over forty locations in the basin ranging in elevation from just above the modern lake to just below the Bonneville shoreline. More generally, similar deposits are likely to exist in protected spaces in other terminal lake basins, offering the potential for a new generation of precisely dated records of past water balance changes.

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Appendix A. Supporting information

Supplementary data associated with this article can be found in the online version at <http://dx.doi.org/10.1016/j.epsl.2012.07.019>.

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