APPLYING URANIUM-SERIES ISOTOPE
GEOCHEMISTRY AND GEOCHRONOLOGY TO
GREAT BASIN PLEISTOCENE PALEOHYDROLOGY

A THESIS
SUBMITTED TO THE DEPARTMENT OF GEOLOGICAL AND
ENVIRONMENTAL SCIENCES
AND THE COMMITTEE ON GRADUATE STUDIES
OF STANFORD UNIVERSITY
IN PARTIAL FULFILLMENT OF THE REQUIREMENTS
FOR THE DEGREE OF
MASTER OF SCIENCE

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December 2013

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ABSTRACT

The widespread lake systems of the Basin and Range during the late Pleistocene indicate substantially greater moisture availability during glacial periods relative to modern. To determine the hydrography of the most recent lake cycle, we dated shoreline tufa deposits from wave-cut lake terraces in Surprise Valley, California. The lake hydrograph is constrained by ^{230}Th -U ages on 22 tufa samples paired with 15 radiocarbon ages. This new lake hydrograph places the highest lake level 176 m above the present-day playa at $>15.23 \pm 0.36$ ka cal BP (^{14}C age). During the Last Glacial Maximum (LGM, ~ 19 to 26 ka), Lake Surprise stood at moderate levels, 65 to 99 m above modern playa. Temporally, the Lake Surprise highstand slightly postdates the Lake Lahontan highstand and corresponds to several post-LGM highstands and stillstands of smaller lake systems farther east.

To further evaluate climatic forcings associated with lake-level changes, we use an oxygen isotope mass balance model combined with an analysis of predictions from the Paleoclimate Model Intercomparison Project 3 (PMIP3) climate model ensemble. Our isotope mass balance model predicts minimal precipitation increases of only 2.5 to 18.2% (average = 9.5%) during the LGM relative to modern, compared to an approximately 75% increase in precipitation during the 15.23 ka highstand when lake surface area increased by 138%. LGM PMIP3 climate model simulations corroborate these findings, predicting an average precipitation increase of only 6.5% relative to modern, accompanied by a 28% decrease in total evaporation propelled by a 7°C decrease in mean annual temperature. LGM climate model simulations also suggest a seasonal decoupling of runoff and precipitation, with peak runoff shifting to the late spring. Based on our coupled analysis, we propose that moderate lake levels during the LGM were driven by reduced evaporation, a result of reduced summer insolation, and not by increased precipitation. Reduced evaporation primed Basin and Range lake systems, particularly smaller, isolated basins such as Surprise Valley, to rapidly respond to increased precipitation during late-Heinrich Stadial 1 (HS1, ~ 14.5 to 19 ka). Post-LGM highstands were potentially driven by increased rainfall during HS1 brought by latitudinally extensive and strengthened mid-latitude westerly storm tracks, the effects of which are recorded in the lacustrine and glacial records as far south as $\sim 32^\circ\text{N}$. These results suggest

that seasonal insolation, in particular the effect of summer insolation on lake evaporation, provides a previously under-investigated long-term driver of moisture availability in the western United States.

ACKNOWLEDGEMENTS

Over the last several years I have been fortunate to have a number of extremely dedicated and passionate advisors and mentors here at Stanford. First and foremost I wish to thank my advisor, Professor Kate Maher, for the support and confidence she has had in me for the last several years. Challenging me to delve deeply into the fields of isotope geochemistry and paleoclimate, Kate always pushes me to maintain high standards in my data collection, writing, and presentations. I am grateful for the mentorship, both academic and personal, that Kate has given me since I was an undergraduate at Stanford, and look forward to continued collaboration with Kate.

Caroline Harris, whom I apprenticed under for several quarters as the ICP-MS/TIMS laboratory's student research assistant, was incredibly even-keeled and instructive while initiating me into the fascinating world of mass spectrometry, no matter how many nebulizers I broke! Karrie Weaver, our clean lab manager, demonstrated patience and persistence in helping me learn the skills necessary to carry out various aspects of my project. Caroline and Karrie, along with fellow student Kimberly Lau, were instrumental in the successful development of the ^{230}Th -U geochronology methods in our lab. Additionally, they both helped me drive forward and maintain momentum during periods of sample processing and writing.

Anne Egger encouraged me to pursue this project further following my senior thesis working on Surprise Valley, and has been an excellent editor to ensure that my paper was intelligible to geologists, and not just paleoclimatologists and isotope geochemists. As my second reader Jorge Vazquez provided valuable reviews, and paid special attention to the methodological details and accurate descriptions of the isotopic analysis portions of my thesis. Jeremy Caves provided valuable editing help and constantly reminded me to step back from my work to think about the broader impacts of the results and interpretations. I am also very appreciative of Perach Nuriel, Matt Winnick, Dan Horton, Jessica Oster, Sverre Leroy, Hari Mix, Miguel Cruz, Conni De Masi, Claire Kouba, Pablo Garcia del Real and Kimberly Lau for taking the time to discuss my work, help with various analyses, and suggest avenues for additional inquiry. I also must express gratitude to Richard Nevle for being a supportive mentor and advocate.

Through several different teaching assistantships (Senior Seminar and Gold Rush), I learnt from Dennis Bird and Simon Klemperer that effective communication of scientific concepts, in both the oral and written formats, is of utmost importance for communicating meaningful and impactful research. I am grateful for their lessons in communicating science, and hope to continue to build on the strong foundation they have provided for me.

I would like to thank all of the students and postdocs in the Maher group, past and present, for supporting me through many successes and failures in the past few years. In particular, my office mates, Dana Thomas, Cynthia McClain and Miguel Cruz, along with Valerie Rosen, Pablo Garcia del Real, Kimberly Lau and Joey Nelson, have created an intellectually stimulating and supportive environment to pursue research. My friends at Stanford, the Sunday evening dinner crew, provided a community that has supported me unconditionally. Zach Ming, Jason Gonzalez, Kaz Gunning, Krystina Tran, Julia Brownell, Vincent Chen, Erin Olivella-Wright and most importantly Sarah Lummis, are all inspirational and talented individuals who have been a major source of encouragement in my daily life.

My parents, Diana and Dennis, and my brother, Eric, have been faithful supporters of me throughout my education. My parents desire to foster my exploration and appreciation of the natural world during my childhood led to numerous family trips to national parks and home chemistry experiments that undoubtedly foreshadowed my foray into geochemistry. My parents have always encouraged me to follow my dreams and pursue my passions, and I am grateful for their unwavering support.

Analytical measurements and fieldwork expenses were supported by NSF EAR-0921134 award to Kate Maher. Lidar data was collected by National Center for Airborne Laser Mapping (NCALM) with funding from NASA through award 10-UAS10-0021 to Jonathan Glen, Anne Egger, and Corey Ippolito. When not working as a teaching assistant or research assistant I was partially supported by the Thomas D. & Janice H. Barrow Fellowship from the Department of Geological & Environmental Sciences.

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Preface

The principle work presented here, focusing on Surprise Valley, California, continues recent work by other researchers and students from Stanford University in the northwestern margin of the Basin and Range. The research groups led by Professor Elizabeth Miller and Professor Simon Klemperer, in collaboration with USGS scientists Jonathan Glen and David Ponce, have predominantly spearheaded geologic and geophysical work in the Surprise Valley region. Joe Colgan, Derek Lerch, Matt Coble, Anne Egger, Valentina Fontiveros and Noah Athens have all completed dissertations and theses working on various aspects of the Cenozoic volcanic and extensional history of Surprise Valley and the northwest Great Basin region (Colgan, 2005; Lerch, 2007; Egger, 2010; Fontiveros, 2010; Athens, 2011; Coble 2012). This thesis is the first to focus on surface geochemistry, hydrology and Quaternary geology. Building on the mapping of Anne Egger (Egger and Miller, 2011), Quaternary fault mapping by Charles Hedel (1980; 1984), and work by the USGS on Quaternary slip along the Surprise Valley Fault (Personius et al. 2009) this work is the first work to decipher the lacustrine Quaternary geology of Surprise Valley and quantify the latest Pleistocene lake level fluctuation of Lake Surprise. My senior thesis (Ibarra, 2012), *An assessment of late Pleistocene to middle Holocene lake level fluctuations in Surprise Valley, California*, served as reconnaissance for the larger body of work presented in this masters thesis. Following this preface is a list of scientific contributions resulting from this thesis.

The main body of the thesis covers the latest Pleistocene history of Lake Surprise, California. This manuscript, *The latest Pleistocene history of Lake Surprise, California: Quantifying controls on the water cycle during the last deglaciation*, has been submitted to *Geological Society of America Bulletin*. Appendix A presents the data repository figures and tables submitted to *Geological Society of America Bulletin*. This publication will be co-authored by Anne Egger, Karrie Weaver, Caroline Harris and Kate Maher. Appendix B is an extended description of the uranium-series geochronology methods developed in conjunction with this project, and other related research projects (Oster et al., 2012; Maher et al., *In Press*) in collaboration with Caroline Harris, Karrie Weaver, Jessica Oster, Kimberly Lau, Miguel Cruz and Kate Maher. Appendix C provides

additional discussion of the methods used to process the Paleoclimate Model Intercomparison Project 3 (PMIP3) climate models.

Scientific contributions resulting from this thesis

Publications (peer-reviewed):

Ibarra, D.E., Egger, A.E., Weaver, K.L., Harris, C.R., Maher, K., *In Review*, The latest Pleistocene history of Lake Surprise, California: quantifying controls on the water cycle during the last deglaciation. Submitted to *Geological Society of America Bulletin*.

Conference abstracts:

Ibarra, D.E., Weaver, K.L., Harris, C.R., Maher, K., 2013, A combined U-series, radiocarbon and stable isotope approach for constructing a Pleistocene lake hydrograph: an example from Surprise Valley, California. Abstract PP13B-1879 presented at the 2013 Fall Meeting, AGU, San Francisco, California, 9-13 December.

Ibarra, D.E., Egger, A.E., Weaver, K.L., Harris, C.R., Maher, K., 2013, Predicting controls on the water cycle in the northwestern Great Basin during the last deglaciation: Surprise Valley, California. Oral presentation 308-10 presented at the 2013 Annual Meeting, GSA, Denver, Colorado, 27-30 October. Geological Society of America Abstracts with Programs, v. 45, no. 7.

Ibarra, D.E., Egger, A.E., and Maher, K., 2013, The late Pleistocene pluvial history of Surprise Valley, CA. Poster presented at the 26th Pacific Climate Workshop. Pacific Grove, California, 3-6 March.

Ibarra, D.E., Maher, K., Oster, J.L., Egger, A.E., Harris, C.R., Horton, D.E., Weaver, K.L., 2012, Comparing lake and soil records to climate model simulations of hydrologic conditions across the western United States at the Last Glacial

Maximum. Oral presentation PP4A-07 presented at the 2012 Fall Meeting, AGU, San Francisco, California, 3-7 December.

Egger, A.E., and **Ibarra, D.E.**, 2012, Paleoseismology from paleoshorelines: Combining lidar data and geochronology to resolve displacement of Pleistocene pluvial shorelines in the northwestern Basin and Range. Abstract PP11A-2003 presented at the 2012 Fall Meeting, AGU, San Francisco, California, 3-7 December.

Ibarra, D.E., Maher, K., Egger, A., Oster, J.L., Harris, C.R., and Weaver, K., 2012, Combining soil and lake records with AOGCM simulations to assess hydrologic conditions in western North America at the Last Glacial Maximum. Oral presentation at the 2012 6th Graduate Climate Conference, Pack Forest Conference Center, University of Washington, 26-28 October.

Research presentations:

Ibarra, D.E., Egger, A.E., and Maher, K., 2013, The late Pleistocene pluvial history of Surprise Valley, CA. Poster presented at the Stanford University School of Earth Sciences Research Review, 12 April.

Ibarra, D.E., 2012, Enhancing paleo-lake level records of an internally draining watershed: Surprise Valley, California, Geographic Information Systems (GIS) Day, Branner Earth Sciences Library, Stanford University, 14 November. (Invited)

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The latest Pleistocene history of Lake Surprise, California: quantifying controls on the water cycle during the last deglaciation

1. INTRODUCTION

The late Pleistocene landscape of the western United States was characterized by vast lake systems indicative of a hydrologic balance dramatically altered from the present (**Figure 1**; Mifflin and Wheat, 1979; Reheis, 1999a). However, uncertainty in the timing of major hydrologic changes has made it difficult to connect the observational record to well-dated climatic events. In addition, the precise connection between lake levels and climate forcings has proven challenging to establish, because the relationships among the physical and hydrologic controls on measured variables and past climatic states are unresolved. This study, a new lacustrine paleoclimate record from Surprise Valley, California, reconciles how different forcings, namely the competing influences of solar insolation and increased precipitation due to atmospheric teleconnections, influenced Pleistocene lakes in the Basin and Range during the last deglaciation.

The geographic extent and temporal trends of the latest Pleistocene lake levels require that orbital conditions and changes in atmospheric circulation impose wetter and/or cooler conditions on the western United States. In the Basin and Range, the majority of lake highstands occurred between 15 to 18 ka, during Heinrich Stadial 1 (HS1, ~14.5 to 19 ka), several thousand years after the Last Glacial Maximum (LGM, ~19 to 26 ka) (e.g., Benson et al., 1990; Adams and Wesnousky, 1998; Garcia and Stokes, 2006; Munroe and Laabs, 2012). These lakes appear to have stood at moderate water levels during much of the early Marine Oxygen Isotope Stage 2 (MIS 2, ~11 to 29 ka) (e.g., Benson et al., 1995; Wells et al., 2003; Bacon et al., 2006), and low to moderate levels through MIS 3 (~29 to 57 ka) (e.g., Tackman, 1993; Phillips et al., 1994; Reheis et al., 2012). Recent syntheses of lake highstands 31-43°N (e.g., Lyle et al., 2012; Munroe and Laabs, 2012) suggest a temporal correspondence between lake highstands and HS1. The atmospheric mechanism driving these high lake levels during HS1 is hypothesized to be mid-latitude ‘dipping westerlies’ (Negrini, 2002), reaching as far south as Lake Elsinore, California (~34°N, Kirby et al., 2013) and Cave of the Bells, Arizona (~32°N, Wagner et al., 2010). Despite evidence from late Pleistocene lake records and other

paleoclimate archives, the temporal correspondence between (Munroe and Laabs, 2012) or robust latitudinal trends in (Lyle et al., 2012) lake highstands and stillstands during HS1 are unresolved. Knowledge of the spatial distribution of hydrologic shifts, recorded by paleoclimate archives, is required to resolve the underlying climatic drivers of profound hydrologic change in the late Pleistocene. Furthermore, the mechanisms (e.g. reduced temperatures and lake surface evaporation, increased rainfall) that produced moderate lake levels during the LGM, before the deglacial highstands, are not well understood.

Most studies infer Pleistocene lake levels to record precipitation amounts driven by changes in mid-latitude atmospheric circulation. A large range of estimates for the extent of increased precipitation (80 to 260% of modern), reduced evaporation (12 to 90% of modern) and decreased temperature (3 to 15 °C lower than modern) during the LGM have been calculated using a multitude of models (cf. Matsubara and Howard, 2009, Table 1). These broad ranges of estimates pose a problem for Atmosphere-Ocean Global Climate Models (AOGCMs), which require spatially resolved changes in the hydrologic cycle for data-model intercomparison (e.g., Braconnot et al., 2007, 2012; DiNezio and Tierney, 2013). As a supplement to well-dated late Pleistocene lake hydrographs, stable isotopes provide an additional constraint on watershed-scale moisture availability driven by larger changes in the climate system (e.g. Hostetler and Benson, 1994; Jones et al., 2007; Placzek et al., 2011).

While most research has focused on the large and complex lake systems of Lake Lahontan and Lake Bonneville (**Figure 1**, e.g., Benson et al., 1990; Oviatt et al., 1992; Adams and Wesnousky, 1998; Godsey et al., 2011; McGee et al., 2012), small lakes – which record hydrologic conditions of smaller watersheds – are more responsive to climate fluctuations and thus provide higher-resolution paleoclimatic data (Garcia and Stokes, 2006; Munroe and Laabs, 2013; Steinman et al., 2013). Given a mean climatic state, application of the ‘water equivalency rule’ (Hendriks et al., 2012) to small lake systems implies that they will produce the most valuable paleoclimate records needed to decipher temporal changes in the hydrologic cycle. Thus, lakes with small watersheds, simple hypsometry and short residence times will record perturbations in the hydrologic cycle more rapidly and with less lag.

Given the utility of small lake systems for paleoclimate reconstructions, we examine the latest lake cycle of Surprise Valley in northeastern California (**Figure 1**). This small valley is ideal for testing paleoclimate models because it is located between two major climatic zones. In particular, LGM climate models predict a more arid Pacific Northwest and a relatively wetter central Nevada (Kim et al., 2008; Laîné et al., 2009; Braconnot et al., 2007; 2012). Surprise Valley, in northeastern California, is located within the transition between these two climatic zones. Paleoclimate records from the north and west suggest reduced precipitation and a mean annual temperature $\sim 7^{\circ}\text{C}$ lower during the LGM (Worona and Whitlock, 1995; Bradbury et al., 2004). To the southeast, Lake Lahontan (**Figure 1**) stood at moderate levels during the LGM (Benson et al., 1995), prior to a brief deglacial highstand at ~ 15.8 ka (Adams and Wesnousky, 1998).

We use a ‘shore-based’ approach (e.g., Redwine, 2003; Garcia and Stokes, 2006; Kurth et al., 2011; Munroe and Laabs, 2013) to quantify late Pleistocene lake levels in Surprise Valley. Pleistocene Lake Surprise hydrography is recorded in prominent wave-cut shorelines along the steep valley walls of Surprise Valley. While lake sediment core studies can provide higher resolution climate archives (e.g., Benson et al., 1990; Bischoff et al., 1997a,b; Licciardi, 2001; Rosenbaum et al., 2012), records of shoreline ages document the history of lake surface area, a first order measure of the integrated balance of precipitation and evaporation for a given basin (Mifflin and Wheat, 1979; Benson and Paillet, 1989; Currey, 1990; Reheis, 1999b; Munroe and Laabs, 2013). We incorporate stable isotopes into a simplified hydrologic mass-balance model (rooted in changes in lake surface area), which allows us to quantify the changes in precipitation and evaporation during lake level fluctuations, and provide direct comparison to climate model outputs. To obtain age-elevation constraints on lake level, we dated carbonate (tufa) deposits on wave-cut bedrock using ^{230}Th -U and radiocarbon (^{14}C) geochronology. In addition to filling temporal and geographical gaps in the paleoclimate record of the western United States, we seek to test climate model predictions for the LGM at a location where models diverge in their spatial prediction of past rainfall magnitude and net change in a small basin that does not integrate over a large area.

Our new lake record and isotope mass balance calculations provide evidence to help reconcile three key questions regarding the timing and mechanisms for the latest

Pleistocene lakes in the Basin and Range: (1) Can moderate Last Glacial Maximum (LGM) lake levels be explained by lower evaporation rates due to decreasing summer solar insolation? (2) Is increased precipitation, driven by changes in atmospheric circulation, required to explain post-LGM highstands during late-Heinrich Stadial 1 (HS1)? (3) Is there a temporal correspondence or robust latitudinal trend in lake highstands and stillstands during HS1 that records changes in the position, strength and/or character of mid-latitude atmospheric circulation?

2. BACKGROUND

2.1. Geologic setting and previous work

Surprise Valley, located in the northeast corner of California and the westernmost margin of the Basin and Range (**Figure 1**), is situated within a climatic transition from semi-arid coniferous forest in the west to arid sagebrush- and grassland-dominated high altitude desert in the east. The N-S-trending valley formed as a result of extension since the mid-Miocene, and is bound by normal faults along the range fronts of the Warner Range to west and the smaller Hays Canyon Range to the east (cf. Personius et al., 2009; Egger et al., 2010; Lerch et al., 2010; Egger and Miller, 2011). Playas in the upper, middle and lower sub-basins occupy the valley floor (**Figure 2A**). To the south lies Duck Flat, a small sub-basin hydrologically connected to the Surprise Valley watershed (**Figure 2A**).

Although nearby Lake Lahontan and the Chewaucan Basin have been the subject of detailed study (**Figure 1**), few studies have focused on Pleistocene Lake Surprise. Paleoshorelines in Surprise Valley appear laterally continuous throughout the valley and into Duck Flat (Russell, 1884; Russell, 1927; Hubbs and Miller, 1948). On the east side of Surprise Valley, wave-cut shoreline features are eroded into bedrock, cutting into mid- to late-Cenozoic rhyolite, basalt and tuff (Carmichael et al., 2006; Egger and Miller, 2011; Colgan et al., 2011). Laminated shoreline tufa on exposed bedrock is abundant on several shorelines at elevations of 1420 to 1450 m (all elevations given as meters above sea level (a.s.l.)), but is less common at higher and lower elevations (**Figure 2**). Personius et al. (2009) noted that the latest Pleistocene highstand appears to be found throughout the valley at ~1540 m. Additionally, two late Pleistocene highstand elevations have been

proposed: Mifflin and Wheat (1979) and Reheis (1999a) both reported highstand elevations of 1567 m. In contrast, Irwin and Zimbelman (2012) report a highstand elevation of 1545 m.

Analysis of modern topography by Irwin and Zimbelman (2012) suggests that the basin pour point is 1621 m, significantly higher than any of the reported highstand elevations, indicating that Lake Surprise was a terminal lake. In addition, groundwater quality studies indicate that groundwater flows towards the center of the basin (Clawson et al., 1986). Russell (1927) noted geomorphic evidence that suggests canyon-carving overflow into the Lahontan Basin through Buffalo Meadows during much older and higher highstands similar to other Great Basin lake systems (including Lahontan), likely recording progressive drying from the early middle to late Pleistocene (Reheis, 1999a,b; Reheis et al., 2002; Kurth et al., 2011). However, field and modeling studies investigating pluvial lake surface areas, hydrography, and watershed run-off have assumed that late Pleistocene Lake Surprise was inward draining and was not connected to the much larger Lake Lahontan system to the south or to Warner Valley to the northeast (**Figure 1**, e.g., Russell, 1884; Hubbs and Miller, 1948; Mifflin and Wheat, 1979; Benson and Paillet, 1989; Benson et al., 1995; Adams and Wesnousky, 1998; Sack, 2002; Matsubara and Howard, 2009).

Late Pleistocene hydrographs have been produced for Great Basin lake systems including Lahontan (e.g., Benson et al., 1995; Adams et al., 2008), Franklin (Lillquist, 1994; Munroe and Laabs, 2013), Searles (Smith, 1984), Bonneville (e.g., Oviatt et al. 1992), Russell (e.g., Zimmerman et al., 2011) and Chewaucan (Licciardi, 2001) (**Figure 1**). To date, no similar hydrographs exist for Lake Surprise. Personius et al. (2009) summarized several existing time-depth constraints on Pleistocene hydrography based on the distribution of lacustrine sediments deposited over the Surprise Valley Fault along the western margin of the basin (**Figure 2A**).

We summarize the existing time-depth constraints for Lake Surprise's latest Pleistocene lake cycle below (further summarized in **Table A1**¹). The Trego Hot Spring tephra (THS), with an assigned radiocarbon age of 23.2 ± 0.3 ka cal BP from nearby Pyramid Lake (Benson et al., 1997), is exposed in two locations in the valley (**Figure 2A**; Personius et al., 2009; Hedel, 1980; 1984) at ~1378 m. Hedel (1980) notes that the THS

at both localities is deposited in fine sand and silt, likely associated with moderate to deep lake levels. Additionally, shallow water deltaic deposits at 1475 and 1493 m postdate the THS deposit and likely represent deposition between 18 and ~ 24 ka (Personius et al., 2009). Personius et al. (2009) also report subaerial sediment deposition at 1475 m from 1 to 13 ka (**Figure 2A**). Finally, archaeological evidence from three sites excavated along the edges of the desiccated Holocene playa lakes (with a minimum elevation of 1355m)—King’s Dog, Menlo Baths and Rodriquez (**Figure 2A**)—indicates that, since the middle Holocene, two distinct prehistoric populations, the Modoc and Northern Paiute, occupied Surprise Valley as a seasonal winter home (James, 1983; O’Connell and Inoway, 1994). At the King’s Dog site a radiocarbon age from a bone on the second deepest excavated house floor was dated at $5,640 \pm 155$ yr BP (^{14}C age), the oldest from all three sites (O’Connell and Inoway, 1994).

Figure 1 (previous page). Location map of late Pleistocene lakes (light blue), modern lakes (dark blue) and the extent of major LGM mountainous glaciation (gray) in the Western United States (simplified from Mifflin and Wheat, 1979; Reheis, 1999; Ehlers et al., 2011). The black box delineates the extent of **Figure 2A**. Locations of additional paleoclimate archives compiled in **Figure 10** and discussed within the text include glacier records (red triangles) and pollen records (red circle). Labeled lakes and pollen records are Alvord Basin (AL), Carpenter Lake (CR), Chewaucan Basin (CB), Columbus Lake (CO), Diamond Valley (DV), Fort Rock (FR), Jakes Lake (JL), Lake Bonneville (LB), Lake Clover (CL), Lake Franklin (LF), Lake Lahontan (LL), Lake Russell (LR), Lake Surprise (SV), Little Lake (LL), Mahleur Lake (MH), Newark Valley (NV), Railroad Lake (RR), Spring Lake (SL), Upper Klamath Lake (UKL) and Waring Lake (WL).

Figure 2 (previous page). Location maps for samples collected as part of this study. Sample locations (latitude, longitude and elevation) are listed in **Table 1**. **(A)** Detailed map of the Surprise Valley region including the calculated Lake Surprise watershed (black line) and outlines of Lake Surprise's extent (green, red, and dark blue lines) at key periods, with water sample (blue circles), playa sample (red squares), and weather station (black stars) locations. Lake Surprise outlines were generated by contouring the merged DEM from the USGS National Elevation Dataset using ArcGIS 10.1. For the LGM lake level (1420-1440 m) only 1440 m is shown on this panel. Only streams within the basin's watershed are included, demonstrating that Lake Surprise was an inward draining, closed basin lake system. Also shown are the locations of additional lake level constraints from archaeological sites (yellow stars) (James, 1983; O'Connell and Inoway, 1994), Trego Hot Spring Tephra localities (green diamonds) (Hedel, 1980) and a paleoseismic trench with lacustrine sediments (gray square) (Personius et al., 2009). Black boxes B, C, D and E delineate the areas of additional panels. **(B-E)** Locations where carbonate samples were collected on exposed shoreline ridges (green circles). Group 1 tufa samples were collected from the prominent lower elevation shorelines at 1419 to 1478 m (white circles); group 2 samples were collected from the less-prominent middle elevation shorelines of 1509 to 1531 m (black circles); and group 3 samples were collected from higher elevation shorelines >1542 m (black diamonds). Each sample locality includes the USGS 7.5' quadrangles (1:24,000 scale), overlain with shaded relief and Quaternary units adapted from Egger and Miller (2011).

2.2. Geochronology of lake shorelines

Assembly of a lake hydrograph provides a first-order constraint on the climatic changes required to sustain Lake Surprise within the closed-basin watershed of Surprise Valley during the late Pleistocene. Techniques to assess the relative ages of Pleistocene surfaces have been successfully applied to Basin and Range shorelines and alluvial fans by using a combination of soil-profile development, tephra chronology, radiocarbon chronology, geomorphic position, and the extent of CaCO_3 accumulation (e.g. Harden and Taylor, 1983; Tackman, 1993; Redwine, 2003). However, the optimum age dating technique will provide numerical ages. Absolute age-dating techniques applicable to Quaternary surficial surfaces include radiocarbon geochronology (Oviatt et al., 1992; Benson et al., 1995; Zimmerman et al., 2012), ^{230}Th -U geochronology (Maher et al., 2007; Kurth et al., 2011), electron-spin resonance (Jones et al., 2004), amino acid racemization (Kaufman and Miller, 1992; Kaufman, 2003), luminescence (Prescott and Robinson, 1997; Haskell et al., 1999), and the use of cosmogenic nuclides (such as ^{36}Cl , ^{10}Be , ^{26}Al , ^3He) (Licciardi et al., 2004; Phillips et al., 2009; Kurth et al., 2011).

In this study, we targeted wave-cut lake terraces occurring in Surprise Valley using ^{230}Th -U and radiocarbon geochronology of tufas to provide absolute ages of surfaces. Given recent advances in the precision of ^{230}Th -U dating by MC-ICPMS (Hernández-Mendiola et al., 2011; Shen et al., 2012; Cheng et al., 2013), and the establishment of robust radiocarbon calibration curves (Reimer et al., 2004; 2009; Fairbanks et al., 2005), these two methods provide complementary constraints on the age of recent lacustrine carbonate.

2.2.1. ^{230}Th -U dating of impure lacustrine carbonates

^{230}Th -U dating has been successfully applied to a range of terrestrial carbonate materials, most notably to speleothems (e.g., Vacco et al., 2005; Oster et al., 2009; Wagner et al., 2010; Polyak et al., 2012), pedogenic calcite (e.g., Ludwig and Paces, 2002; Sharp et al., 2003; Fletcher et al., 2010), travertine (e.g., Luo and Ku, 1991; Soligo et al., 2002), and a variety of lacustrine carbonates (e.g., Israelson et al., 1997; Haase-Schramm et al., 2004; Placzek et al., 2006a,b; Blard et al., 2011; McGee et al., 2012; Torfstein et al., 2013). Lacustrine carbonate associated with tufa mounds and shoreline tufa deposits is often found in late Pleistocene lake systems in the western United States

(e.g., Benson et al., 1995; Ku et al., 1998; Godsey et al., 2011; Zimmerman et al., 2011) and elsewhere (e.g., Moeyersons et al., 2006; Placzek et al., 2006a,b; 2011; Haberlah et al., 2010; Blard et al., 2011), making them useful for paleo-lake level reconstructions.

^{230}Th -U dating relies on the first three long-lived daughter products in the ^{238}U decay series. Uranium, a radioactive element, consisting of 99.28 % ^{238}U ($t_{1/2} = 4.46$ Gyr) (Jaffey et al., 1971) decays via the following sequence: $^{238}\text{U} \rightarrow ^{234}\text{Th} \rightarrow ^{234}\text{Pa} \rightarrow ^{234}\text{U} \rightarrow ^{230}\text{Th}$. The ^{234}Th ($t_{1/2} = 24.1$ days) and ^{234}Pa ($t_{1/2} = 1.18$ min) daughter isotopes are very short-lived compared to ^{238}U , ^{234}U and ^{230}Th . Samples can thus be dated using the activity ratios of the two long-lived daughter isotopes ($^{234}\text{U}/^{238}\text{U}$) and ($^{230}\text{Th}/^{238}\text{U}$), where parentheses denote activity ratios. Due to the relatively short half-lives of ^{230}Th ($t_{1/2} = 75.4$ kyr) and ^{234}U ($t_{1/2} = 245$ kyr), ^{230}Th -U geochronology has an upper age limit of about ~500-800 ka depending on analytical precision (Cheng et al., 2013). The equations for calculating the ingrowth of ^{230}Th and the closed-system decay of ^{234}U are (Kaufman and Broecker, 1965; Thurber et al., 1965; Neymark, 2011):

$$\left(\frac{^{234}\text{U}}{^{238}\text{U}}\right)_t = (1 - e^{-\lambda_{234}t}) + \left(\frac{^{234}\text{U}}{^{238}\text{U}}\right)_0 \quad (1)$$

$$\left(\frac{^{230}\text{Th}}{^{238}\text{U}}\right)_t = \left(\frac{\lambda_{230}}{\lambda_{230} - \lambda_{234}}\right) \left(\frac{^{234}\text{U}}{^{238}\text{U}}\right)_0 (e^{-\lambda_{234}t} - e^{-\lambda_{230}t}) + \left(1 + \frac{\lambda_{234}e^{-\lambda_{230}t} - \lambda_{230}e^{-\lambda_{234}t}}{\lambda_{230} - \lambda_{234}}\right) \quad (2)$$

where λ_{230} and λ_{234} are decay constants for ^{230}Th and ^{234}U , respectively, t is the time in years since the mineral formed and $(^{234}\text{U}/^{238}\text{U})_t$ and $(^{230}\text{Th}/^{238}\text{U})_t$ are the measured or corrected activity ratios. These equations are solved for $(^{234}\text{U}/^{238}\text{U})_0$ and time (t). In closed-system rocks and minerals older than $\sim 10^6$ years, the $(^{234}\text{U}/^{238}\text{U})$ and $(^{230}\text{Th}/^{238}\text{U})$ are equal to one, a condition referred to as “secular equilibrium.” The $(^{234}\text{U}/^{238}\text{U})$ of natural water is typically elevated (>1) due to the preferential release of the daughter ^{234}Th nuclide across grain boundaries during the α -decay of ^{238}U , a process known as α -recoil (Kigoshi, 1971; Fleischer, 1982; Hussain and Krishnaswami, 1980; Oster et al., 2012). Under oxidizing conditions typical of surface waters, U is mobile as the hexavalent uranyl ion ($(\text{U(VI)}\text{O}_2)^{2+}$) and is incorporated into secondary minerals such as carbonates (Reeder et al., 2001). In contrast, Th is relatively insoluble under oxic conditions, and ideal samples have low initial Th concentrations (e.g., $(^{230}\text{Th}/^{238}\text{U}) = 0$) such that all post-depositional ^{230}Th accumulation is derived from radioactive decay of

^{234}U in the sample. However, by measuring the abundance of ^{232}Th ($t_{1/2} \sim 14$ Gyr) in the sample, corrections can be made for any “detrital Th” (i.e., Th derived from other processes aside from purely radioactive decay) that may have been originally incorporated into the carbonate (e.g., Ludwig and Titterton, 1994; Neymark and Paces, 2000; Sharp et al., 2003; Paces et al., 2004; Fletcher et al., 2010).

The assumption of no initial ^{230}Th is rarely valid for lake carbonates (e.g., Placzek et al., 2006a,b; Blard et al., 2011; Torfstein et al., 2013), which often contain one to several potential sources of additional Th. In Mono Basin (**Figure 1**, Lake Russell) for example, it has been observed that unsupported Th, which may co-precipitate with lacustrine carbonates, can be the result of high Th activity and complexation with CO_3^{2-} . This inhibits removal and scavenging of hydrogenous Th by absorption to particles (Anderson et al., 1982), as observed in ocean sediments (Bacon and Anderson, 1982). In most cases, without correction for initial ^{230}Th , the measured ($^{230}\text{Th}/^{238}\text{U}$) is too high, resulting in an age calculation that is erroneously old. Lake carbonates are likely to have two potential sources of Th that constitute end-members: (1) hydrogenous Th in the primary carbonate, derived from water soluble Th, and (2) detrital Th incorporated within the carbonate as detrital siliciclastic or organic particles (e.g., Lin et al., (1996), Blard et al., (2011) and Torfstein et al., (2013)). In freshwater systems, low hydrogenous Th concentrations are expected; however, some studies of paleolake systems have observed substantial contribution from both Th end-members (e.g., Lin et al., 1996; Haase-Schramm et al., 2004). Assuming that freshwater lake carbonates have only one end-member initial ($^{230}\text{Th}/^{232}\text{Th}$), likely from the siliciclastic or “detrital” component of the sample, we will apply and assess two complementary approaches, single sample (SS) correction and an isochron approach, to calculate the ^{230}Th -U ages for coeval sets ($n \geq 5$) of whole rock dissolutions.

2.2.2. Radiocarbon dating of lacustrine carbonates

Radiocarbon geochronology has been extensively applied to inorganic and organic carbonates found as surficial deposits (e.g., tufa) and in near-shore sediments to date Pleistocene shorelines (e.g., Oviatt et al., 1992; Lin et al., 1996; Garcia and Stokes, 2006; Munroe and Laabs, 2013). Assuming a carbonate sample contains carbon originally fixed from the atmosphere, radiocarbon geochronology relies upon the decay

of ^{14}C ($t_{1/2} = 5.73$ kyr) to ^{14}N to calculate a radiocarbon (^{14}C) age. Because the production and reservoir of ^{14}C in the atmosphere has not been constant with time, calibration datasets (e.g., Reimer et al., 2009) are required to calculate the absolute age of samples. Beyond the Holocene, uncertainty in radiocarbon ages can be the result of plateaus occurring in the radiocarbon calibration curve during the last deglaciation and the short half-life of ^{14}C (Trumbore, 2000). In lacustrine carbonates, precipitating from lake waters, two complications may occur. A reservoir effect due to the incorporation of significant quantities of ‘dead’ carbon (i.e., low $^{14}\text{C}/^{12}\text{C}$ carbon from draining of carbonate bedrock), results in a ^{14}C date that is up to thousands of years older than the true age. Alternatively, in poorly consolidated, porous or impure carbonates, modern or younger atmospheric carbon can contaminate the sample (Cassata et al., 2010; Zimmerman et al., 2011, 2012). In an ideal situation, application of several absolute dating methods, such as ^{230}Th -U or $^{40}\text{Ar}/^{39}\text{Ar}$, applied to coeval or depositionally equivalent material can constrain the reservoir effect or identify anomalously young ages (e.g., Cassata et al., 2010; Zimmerman et al., 2011; Vazquez and Lidzbarski, 2012).

3. METHODS

3.1. Topographic analyses

3.1.1. *Calculation of Lake Surprise basin geometry*

Building on recent work by Irwin and Zimbelman (2012), we used modern Surprise Valley topography to calculate lake volume and surface area at elevations spanning the modern playa to 10 m above the proposed highstand of Reheis (1999). Six 10 m digital elevation model (DEM) datasets from the USGS National Elevation Dataset (NED) were merged, projected to preserve area, and analyzed using the ArcGIS 10.1 hydrology tools and 3-D spatial analyst tools to conduct raster analyses. Using the modern topography, lake surface area and volume were calculated from the minimum playa elevation to above the maximum lake terrace elevation (1355 to 1577 m). To calculate volume, a triangulated irregular network (TIN) surface representation was created using the DEM raster and intercepted with lake surface area. Empirical relationships, fit using polynomial functions, were established to relate volume and surface area to elevation (**Figure A2**). The inward draining watershed area was calculated

after smoothing the DEM. We assume that modern contours and the basin topography accurately represent the watershed topography and lake hypsometry of the late Pleistocene, with minimal isostatic rebound. The results of the Lake Surprise basin geometry analysis were also compared to work by previous authors (Mifflin and Wheat, 1979; Reheis, 1999; Zimbelman et al., 2009; Irwin and Zimbelman, 2012).

Based on analyses of the high-resolution topographic dataset, we updated topographic results from Irwin and Zimbelman (2012) (**Table 2**). The pour point elevation (1621 m) is >50 m higher than all proposed highstand elevations (Hubbs and Miller, 1948; Mifflin and Wheat, 1979; Reheis et al., 1999a; Personius et al., 2009; Zimbelman et al., 2009; Irwin and Zimbelman, 2012), and the pre-MIS 2 and MIS 2 highstands studied here. Furthermore, all modern streams drain terminally into the playa lakes (**Figure 2A**). Groundwater contribution to Lake Surprise is assumed to be negligible as the Plio-Pleistocene lacustrine sediments provide the primary groundwater storage capacity for modern day Surprise Valley and are recharged by near-shore lacustrine deposits and Holocene alluvial fans (Clawson et al., 1986). The groundwater aquifer capacity is estimated to be 4.93 km³ (4 million acre-feet) to 122 m below the valley floor (Clawson et al., 1986), which is only 2.9 % of the lake volume at its latest Pleistocene highstand of 1531 m (lake volume = 164 km³, **Table 2**).

3.2. Field methods

3.2.1. Modern water and playa samples

Modern water samples were collected from seven streams, five groundwater wells, six hot springs and one small pond within the Surprise Valley watershed (**Table 1**; **Figure 2A**). The water samples were filtered in the field through 0.45 µm polyethylene filters into 1-liter polyethylene bottles. The water samples were acidified with high-purity nitric acid (HNO₃) to ~2 % HNO₃. Ten ml were aliquoted for concentration analyses (U, Ca and Sr) and 250 to 500 ml were dried down over ~2 weeks in 50 ml Teflon beakers for U isotope analysis.

Four modern playa surface sediment samples were collected from the playas found in the Surprise Valley watershed (**Table 1**, **Figure 2A**). One modern playa

evaporite sample was collected from the middle playa. All water and playa samples were collected in July 2012.

3.2.2. Tufa sample collection

Tufa samples were collected from exposed bedrock on the front edge or the crest of wave-cut lake terraces and horizontal shoreline benches. Erosional shoreline sets in Surprise Valley are similar to those found in Lahontan (e.g., Adams and Wesnousky, 1998) and Bonneville (e.g., Schofield et al., 2004; Felton et al., 2006; Jewell, 2007). Four sample localities were targeted on the east side of Surprise Valley (**Figure 2**) adjacent to each of the three modern playa lakes (lower, middle and upper alkali lakes). Apparent shoreline sets were targeted on the west side of the Valley, but minimal suitable *in situ* material was found. The east side of the valley has the prominent wave-cut shoreline features, likely because of the prevailing westerly winds. Tufa samples were collected in August 2011 and July 2012. While the absolute depth of tufa formation relative to lake level is not certain or directly quantifiable, their association with wave-cut shoreline features and biologic textures (cf. Rouchy et al., 1996) indicate their formation within the photic zone and likely near the lake surface and edge (see **Table 1** and **Figure 2B-E** for tufa sample locations). Thus, the tufa sample elevations give minimum lake surface elevations at their time of deposition (e.g., Felton et al., 2006; Benson et al., 1996; 2011; Zimmerman et al., 2012).

Latitude and longitude of sample locations were recorded with a handheld GPS (Garmin Oregon 550t). Using a newly acquired lidar dataset, sample elevations were determined by pinning the location coordinates to the 0.5 m horizontal resolution raster digital elevation model derived from the lidar point cloud. The elevation error associated with the elevation of the tufa samples, primarily collected from exposed wavecut shorelines, is ± 10 cm. The location and elevation of all samples are located in **Table 1** and **Figure 2**.

3.3. Analytical procedures

The following is a summary of the laboratory work carried out to determine the Lake Surprise hydrograph during the last glacial cycle, including the application of ^{230}Th -U geochronology, stable isotopic analyses, elemental analyses and accelerator mass

spectrometer (AMS) radiocarbon (^{14}C) geochronology on tufa samples. To support the ^{230}Th -U geochronology, we also analyzed the elemental and U isotopic composition of modern playa sediments (using leachate methods) and modern water samples.

3.3.1. Concentration measurements (Sr, Ca, and U)

3.3.1.1. Modern playa leaches

The range of expected ($^{234}\text{U}/^{238}\text{U}$) in modern playa and water samples is used to improve confidence in our ^{230}Th -U geochronology and assess for potential open-system behavior. An authigenic CaCO_3 (and exchangeable fraction) was extracted based on a sequential leach protocol established by Tessier et al. (1979). These leaches on surface playa sediments from the four sub-basins of Surprise Valley (the upper, middle, lower and Duck Flat playas) were used to determine the ($^{234}\text{U}/^{238}\text{U}$) of the playa sediment components. The leach procedure follows previously established methods (Maher et al., 2003, 2006; DePaolo et al., 2006; Oster et al., 2012):

1. *Pore water*: 1:1 dry playa sediments to deionized water leaches were prepared using 25-30 g of dry sediment and 25-30 ml 18 M Ω deionized water in polyethylene centrifuge tubes. After 24 h, the mixtures were vacuum filtered through 0.22 μm polyethylene filters into screw top polyethylene centrifuge tubes. It has been suggested that deionized water leaches can lead to preferential leaching of U from mineral surfaces on longer (many days to months) timescales (Bonotto et al., 2001). However, due to the short timescales over which these sediments were exposed to water, the ions in these rinse waters are assumed to be only those that were originally dissolved in the pore fluid, and any readily exchangeable ions from mineral surfaces.
2. *Authigenic CaCO_3 (and exchangeable fraction)*: The carbonate fraction of the sediment was extracted using a solution of 1 M sodium acetate (NaOAc). Four grams of soil were mixed with 7.5 ml of 1M NaOAc in acetic acid (buffered to pH 5) following pore water extraction. Samples were allowed to sit for ~ 2 h, after which they were centrifuged, and the supernatant was decanted and saved. The samples were washed twice with 10 ml of 18 M Ω deionized water and centrifuged. The wash solutions were added to the NaOAc leachates and dried

down prior to chemical separation of the uranium to measure the ($^{234}\text{U}/^{238}\text{U}$) of the authigenic CaCO_3 .

3.3.1.2. Digestion of carbonates for elemental analyses

Following the procedures outlined by Davis et al. (2009), 50-100 mg of crushed sample were digested in 14 N nitric acid (HNO_3), diluted with 10 mL of 2 % HNO_3 , sonicated, centrifuged and filtered through 0.45 μm polyethylene filters. Following filtration, 750 μL was further diluted, by adding 6.75 mL of 2 % HNO_3 (corresponding to a 10 to 1 dilution), to measure Sr and Ca concentrations.

3.3.1.3. U and Th concentrations

To determine ^{238}U concentrations of the modern water, 1.5 ml acidified samples were analyzed using a Nu Attom High-Resolution ICP-MS at the Stanford School of Earth Sciences ICP-MS/TIMS facility. Ten concentration standards ranging from 6 to 0.006 ng/ml were prepared gravimetrically and used to build a calibration curve. Standards were run every two samples to monitor and correct for instrumental drift. The concentrations between these standards and samples were matched as closely as possible and washout between measurements was monitored until signals on the electron multiplier returned to background levels. Total gravimetric and analytical error indicates that total standard error (2σ) was 4.3 % for ^{238}U concentrations. Thorium-232 concentrations in the modern surface (stream/lake) water samples were ≤ 6 pg/ml, the apparent detection limit.

3.3.1.4. Sr and Ca concentrations

Sr and Ca concentrations in carbonates and modern waters were measured on a Thermo Scientific ICAP 6300 ICP-OES equipped with a Solid State CID Detector at the Environmental Measurements 1 Laboratory at Stanford University. Total dissolved Ca and Sr were measured at wavelengths 316.9 nm and 346.4 nm / 421.5 nm respectively. The Sr concentrations were taken as the average of both wavelengths. The 421.5 nm concentrations are 99.88 ± 0.10 % of the 346.4 nm concentrations, with high correlation among samples and standards ($R^2 = 0.99$). The error associated with sample preparation blanks (Ca, $2\sigma = 0.17$ %; Sr, $2\sigma = 0.40$ %), and analytical error derived from known gravimetrically prepared standard solutions (Ca, $2\sigma = 2.05$ %; Sr $2\sigma = 1.75$ %) indicate

that total standard error (2σ) was 2.73 % for the Sr/Ca ratios. Procedural and analytical blanks were subtracted from the concentrations.

3.3.2. Stable isotope analysis

Tufa samples and the modern playa carbonate $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ were measured against a known carbonate standard, Vienna PeeDee Belemnite (VPDB). Measurements were made at the Stable Isotope Biogeochemistry Laboratory at Stanford University. 0.05-0.1 mg of crushed, homogenized sample is reacted with orthophosphoric acid at 70 °C in a Thermo Finnigan Kiel III Carbonate Device and measured on a dual inlet Thermo Finnigan MAT 252 stable isotope mass spectrometer. Typical analytical precision of the instrument is <0.06 ‰ for $\delta^{18}\text{O}$ and <0.03 ‰ for $\delta^{13}\text{C}$ analyses. Sample duplicates demonstrated reproducibility < 0.3 ‰ for $\delta^{18}\text{O}$ and < 0.1 ‰ for $\delta^{13}\text{C}$.

3.3.3. Uranium-series isotopic analyses

3.3.3.1. Carbonate sample dissolution

Tufa samples were cut into 0.5-3 gram chips using a microsaw. Dense laminar samples were targeted to avoid areas of high porosity or possible re-crystallization. To remove dust and clean the tufa, each sample was rinsed and sonicated in ethanol.

The clean samples, selected under an optical microscope, were combined with a calibrated mixed spike of ^{236}U and ^{229}Th (calibration details described below), and then dissolved in 14N HNO_3 and treated with hydrogen peroxide (H_2O_2) to remove detrital organic material. Samples that did not fully dissolve contained an insoluble fraction. The nitric acid soluble fraction was decanted and the insoluble fraction was dissolved using aqua regia ($\text{HNO}_3\text{-HCl}$), a concentrated hydrofluoric acid (HF) – nitric acid (HNO_3) digestion, and finally boric acid (H_3BO_3). In preparation for chemical purification, the soluble and insoluble fractions were re-combined, evaporated to incipient dryness, and brought up in 10 ml of 3N HNO_3 for column chemistry.

3.3.3.2. Column Chemistry

U and Th from playa leachates, modern water samples, modern carbonates and spiked Pleistocene tufa samples were purified by ion exchange chromatography by one of two methods. Initially Eichrom TRU-spec resin in 0.2 ml Teflon columns with a 2 ml reservoir was used to separate U and Th following the procedure of Luo et al. (1997). To improve Th yields and separation of Th from U, we switched to an alternate recipe using

Eichrom UTEVA resin modified from Potter et al. (2005) and Stirling et al. (2007). For the playa leachates and modern water samples, only U was collected during column chemistry for isotopic analysis. The UTEVA column chemistry procedure was found to have yields of >95 % for U and >80 % for Th.

Procedural blanks, including dissolution steps, modern water evaporation blanks and column blanks were measured using the Nu Attom ICP-MS. Tufa sample procedural blanks were <3.3 pg of U and <15.3 pg of Th, and column blank blanks contributed <2.0 pg of U and <5.7 pg of Th. Blanks for drying down the 500 ml modern waters samples, over ~2 weeks in 10 ml Teflon beakers, were <16.5 pg of U.

3.3.3.3. Isotopic Analyses

The U and Th column fractions were evaporated to dryness and re-dissolved in 2 % HNO₃ and 2 % HNO₃ + trace HF, respectively. The U and Th isotopic compositions were analyzed separately at the ICP-MS/TIMS Facility at Stanford University, using a Nu Instruments Plasma HR multicollector inductively coupled plasma mass spectrometer (MC-ICP-MS) equipped with 12 Faradays and 4 discrete dynode secondary electron multipliers (ion counters). One of the ion counters (IC0) has an additional decelerating lens filter to improve abundance sensitivity. Measurements were made in dry plasma mode using an Aridus II desolvating nebulizer or a Nu Instrument DSN-100 desolvating nebulizer. All U and Th isotopic standards and solutions were measured in 2 % HNO₃ and 2 % HNO₃ + 0.1 % HF, respectively. Concentrations between samples and standards were matched as closely as possible and washout between measurements was monitored until signals on all detectors returned to background levels.

After gravimetric dilution of a previously calibrated ²²⁹Th/²³⁶U spike, the resulting mixture was re-calibrated by spiking several gravimetrically prepared aliquots of both CRM 145 (n=3) and Th Ames Metal (n=2) at multiple ²³⁶U/²³⁸U and ²²⁹Th/²³²Th ratios. Over the course of sample analysis, the spiked solutions were repeatedly measured as outlined below. The unweighted average concentrations of the newly diluted mixed spike are [²³⁶U] = 0.9618 ± 24 ng/g (n=54) and [²²⁹Th] = 0.2232 ± 20 ng/g (n=58). This ²²⁹Th/²³⁶U ratio of 0.2392 ± 22 mol/mol is in agreement with the pre-dilution calibration of 0.2395 ± 3 mol/mol. Tufa samples were spiked gravimetrically to optimize for a ²³⁶U/²³⁴U ratio of ~5 assuming (²³⁴U/²³⁸U) ≈ 1.5.

U isotopes were analyzed statically with ^{236}U and ^{234}U measured simultaneously on the ion counters IC0 and IC1, and ^{235}U and ^{238}U on Faraday collectors operating with $10^{11} \Omega$ resistors. Voltages on ^{235}U were kept at > 50 mV in order to ensure optimal signal-to-noise ratios on the ^{235}U collector. The measured $^{235}\text{U}/^{238}\text{U}$ ratio was used to correct for instrumental mass fractionation based on an exponential law, and bracketing standards of CRM U010 were used to correct the $^{234}\text{U}/^{238}\text{U}$ and $^{236}\text{U}/^{238}\text{U}$ ratios for Faraday/IC gain. An isotopic standard, CRM 145 (also known as NBL-112A) was analyzed repeatedly during the analytical sessions to assess reproducibility. During the time period of the analyses and spike calibration, the measured $^{234}\text{U}/^{238}\text{U}$ of CRM 145 was 0.000052841 ± 188 (2σ , $n=86$), within error of the certified $^{234}\text{U}/^{238}\text{U}$ of CRM 145 (0.000052841 ± 82 (2σ)). Additionally, an activity standard, SRM 4321A, was measured repeatedly during analytical sessions, giving a $^{234}\text{U}/^{238}\text{U}$ of 0.00005281 ± 19 (2σ , $n=9$), a value in agreement with Oster et al. (2012) (0.00005288 ± 17 (2σ , $n=3$)) and Neymark (2011) (0.00005300 ± 10 (2σ , $n=36$)). Given the accuracy of these standards, no further normalization was required for $^{234}\text{U}/^{238}\text{U}$ measurements.

The Th isotopes were measured dynamically with ^{229}Th and ^{230}Th in IC0 and ^{232}Th in Faraday collectors. Bracketing standards of IRMM-035 were used to correct samples for instrumental mass fractionation and Faraday/IC gain. The Sims et al. (2008) working group $^{230}\text{Th}/^{232}\text{Th}$ average value ($^{230}\text{Th}/^{232}\text{Th} = 0.000011380 \pm 101$ (2σ)) of IRMM-035 was used for normalization. The tail contribution from ^{232}Th onto ^{230}Th was calculated and subtracted from the ^{230}Th signal by concurrently measuring the signal at 0.5 and 0.4 amu on the high mass side of the ^{230}Th peak center and 0.5 amu on the low mass side of the ^{230}Th peak. By scanning the tail across the mass spectrum from <229.5 to >230.5 amu we found that a logarithmic correction provided the best model for the tailing of ^{232}Th onto ^{230}Th (cf. Hoffmann et al., 2007; Sims et al., 2008) and thus a logarithmic tail correction was used to correct standards and samples for the tail contribution at ^{230}Th . The abundance sensitivity at 2 amu was less than 80 ppb for most analyses resulting in a small correction for tailing of ^{232}Th onto ^{230}Th . Typically, the tail correction was between 0.5 % and 1.5 % of the total ^{230}Th signal. The $^{229}\text{Th}/^{232}\text{Th}$ was corrected for gain and mass fractionation using the $^{230}\text{Th}/^{232}\text{Th}$ measurement of the bracketing standard, assuming equal mass bias between $^{229}\text{Th}/^{232}\text{Th}$ and $^{230}\text{Th}/^{232}\text{Th}$.

Isotopic standards IRMM-036 and OU Th “U” were analyzed during the analytical sessions. During the time period of sample analyses the measured $^{230}\text{Th}/^{232}\text{Th}$ for IRMM-036 was 0.000003057 ± 28 ($n = 32$) and for OU Th “U” was 0.000006163 ± 69 ($n = 5$). This is in agreement with the average working group values from Sims et al. (2008), in which the $^{230}\text{Th}/^{232}\text{Th}$ of IRMM-036 is 0.000003047 ± 24 (2σ) and OU Th “U” is 0.000006187 ± 106 (2σ), and other previously published values (Turner et al., 1997; 2001; Thomas et al., 1999; Neymark, 2011; Raptis et al., 1998).

The isotopic measurements (spiked) of a secular equilibrium opal standard, BZVV, from Virgin Valley, NV, yielded average activity ratios of $(^{234}\text{U}/^{238}\text{U}) = 0.998 \pm 0.006$ and $(^{230}\text{Th}/^{238}\text{U}) = 0.989 \pm 0.007$ ($n=4$), in agreement with Amelin and Back (2006). In addition, two basalt rock references of different U and Th concentrations, BCR-2 (USGS reference material from the Bridal Veil Flow Quarry, OR, $n=3$) and TML (Table Mountain Latite from Sonora, CA, $n=2$) were spiked and dissolved in HF-HNO₃. The rock standards were then treated with HNO₃-HCl, dried down, and processed using the modified UTEVA column chemistry described previously. The measured isotope ratios of TML were $(^{230}\text{Th}/^{232}\text{Th}) = 1.074 \pm 0.011$, $(^{234}\text{U}/^{238}\text{U}) = 1.000 \pm 0.002$ and $(^{230}\text{Th}/^{238}\text{U}) = 0.991 \pm 0.009$ with concentrations of $[\text{Th}] = 30.43 \pm 0.72$ ng/g and $[\text{U}] = 10.79 \pm 0.26$ ng/g. The measured isotope ratios of BCR-2 were $(^{230}\text{Th}/^{232}\text{Th}) = 0.855 \pm 0.057$, $(^{234}\text{U}/^{238}\text{U}) = 1.002 \pm 0.003$ and $(^{230}\text{Th}/^{238}\text{U}) = 1.001 \pm 0.083$ with concentrations of $[\text{Th}] = 5.82 \pm 0.68$ ng/g and $[\text{U}] = 1.71 \pm 0.05$ ng/g. Both TML and BCR-2 are in near radioactive equilibrium with respect to the U-series nuclides and agree with literature measurements (Sims et al., 2008, and references within), providing a further verification of both the spike calibration and isotopic measurements described herein. See **Table A2** and **Figure A1** for additional analytical details.

Our results from a variety of reference standards indicate that the methods used for analytical corrections, column chemistry and the calibrated spike concentrations give previously reported and/or certified isotopic values for samples. Isotope ratio errors (internal, external and decay constant errors) were propagated by summation in the quadrature and reported at two sigma significance. ^{230}Th -U ages were determined using methods described in the following section (2-D Rosholt isochrons, 3-D Osmond Isochron and SS age calculation) and calculated using ISOPLOT 3.75 (Ludwig, 2012).

Final activity ratios used the decay constants of Cheng et al. (2013) and Jaffey et al. (1971). Inclusion of decay constant errors in activity ratio error propagation used the “effective errors” of the Cheng et al. (2013) decay constants for ^{234}U and ^{230}Th , intended for ^{230}Th -U dating. 2-D Rosholt isochron ages are reported at the 95% confidence, with errors expanded by a Student's- t multiplier (e.g., Vazquez and Lidzbarski, 2012). 3-D Osmond isochron ages use a first-derivative expansion of the measurement and isochron errors (Ludwig and Tettersington, 1994). The averages of repeat isotope standard analyses noted previously are given as unweighted averages, with results screened at 2σ for standards with >30 analyses (CRM 145 and IRMM-036).

3.3.3.4. *U-series age determination*

To calculate detrital Th corrected ^{230}Th -U ages we relied upon both the isochron approach, involving suites of coeval samples, and SS detrital Th correction. From 21 tufa samples we processed a total of 111 sub-samples. This included 13 sets of 7 coeval samples, two sets of 5 coeval samples from the inner and outer rind of a thick lower elevation sample (SVDI11-T2), and 10 supporting single or duplicate samples. Additionally, to support the construction of isochrons, four 1.5 N HNO_3 leach residues and one modern carbonate sample were also analyzed.

Single sample (SS) correction. For a single (sub-)sample, the ^{230}Th daughter (supported by ^{234}U and ^{238}U) can be determined given an assumed initial ($^{230}\text{Th}/^{232}\text{Th}$) and ^{232}Th concentration (normalized by either ^{234}U or ^{238}U). The measured ($^{230}\text{Th}/^{238}\text{U}$) is corrected using the equation from Kaufman (1993):

$$\left(\frac{^{230}\text{Th}}{^{238}\text{U}}\right)_A = \left(\frac{^{230}\text{Th}}{^{238}\text{U}}\right)_M - \left(\frac{^{232}\text{Th}}{^{238}\text{U}}\right)_M \left(\frac{^{230}\text{Th}}{^{232}\text{Th}}\right)_0 (e^{-\lambda_{230}t}) \quad (3)$$

where the A denotes the authigenic or supported ($^{230}\text{Th}/^{238}\text{U}$), M denotes the measured isotope ratios ($^{230}\text{Th}/^{238}\text{U}$) and ($^{232}\text{Th}/^{238}\text{U}$), 0 denotes the assumed initial ($^{230}\text{Th}/^{232}\text{Th}$) and t is the apparent time calculated before correction for initial Th (cf. Kaufman, 1993; Israelson et al., 1997; Placzek et al., 2006a,b). Once the measured ratio is corrected for initial Th, a corrected age can be calculated via substitution of ($^{230}\text{Th}/^{238}\text{U}$)_A into equation (1). The magnitude and error associated with this correction is a function of the concentration of ^{232}Th in a sample, the assumed ($^{230}\text{Th}/^{232}\text{Th}$)₀ and the error on the assumed ($^{230}\text{Th}/^{232}\text{Th}$)₀. The error on the assumed ($^{230}\text{Th}/^{232}\text{Th}$)₀ is typically the greatest

source of error on the calculated age when $(^{230}\text{Th}/^{232}\text{Th})_{\text{M}} < 10$. While some authors have assumed that $(^{230}\text{Th}/^{232}\text{Th})_0$ is unity (e.g., Sylvestre et al., 1999) or equal to upper continental crust (e.g., Ludwig and Titterton, 1994; Polyak and Asmerom, 2001), it has been observed that in terrestrial aqueous environments, including paleolakes, $(^{230}\text{Th}/^{232}\text{Th})_0$ can vary from 0.5 to 4.2 (Placzek et al., 2006a, Table 5). Thus, for a given system, the most accurate constraint on $(^{230}\text{Th}/^{232}\text{Th})_0$ is via modern lake carbonate measurements (e.g., Lin et al., 1996; Israelson et al., 1997) or an isochron approach (e.g., Placzek et al., 2006a; Blard et al., 2011; Torfstein et al., 2013), as described below. SS ages are calculated for all sub-samples and error-weighted averages were calculated for each set of sub-samples.

Isochron method. A complementary approach to SS corrections is an isochron approach. Isochrons allow for the calculation of the $(^{230}\text{Th}/^{238}\text{U})$ and $(^{234}\text{U}/^{238}\text{U})$ of the pure authigenic carbonate from sets of coeval samples, without necessarily needing to constrain $(^{230}\text{Th}/^{232}\text{Th})_0$. Two mathematically equivalent pairs of 2-D isochrons sharing one common axis or a simultaneous 3-D solution can yield $(^{230}\text{Th}/^{238}\text{U})_{\text{A}}$ and $(^{234}\text{U}/^{238}\text{U})_{\text{A}}$. The 2-D isochrons yield $(^{230}\text{Th}/^{238}\text{U})_{\text{A}}$ and $(^{234}\text{U}/^{238}\text{U})_{\text{A}}$ via the slopes of Rosholt isochrons ($(^{238}\text{U}/^{232}\text{Th})_{\text{M}}$ vs. $(^{230}\text{Th}/^{232}\text{Th})_{\text{M}}$ and $(^{238}\text{U}/^{232}\text{Th})_{\text{M}}$ vs. $(^{234}\text{U}/^{232}\text{Th})_{\text{M}}$) or the y-intercepts of Osmond type isochrons ($(^{232}\text{Th}/^{238}\text{U})_{\text{M}}$ vs. $(^{230}\text{Th}/^{238}\text{U})_{\text{M}}$ and $(^{232}\text{Th}/^{238}\text{U})_{\text{M}}$ vs. $(^{234}\text{U}/^{238}\text{U})_{\text{M}}$) (Osmond et al., 1970; Rosholt, 1976; Luo and Ku, 1991; Bischoff and Fitzpatrick, 1991; Ludwig and Titterton, 1994). Furthermore, the detrital $(^{230}\text{Th}/^{232}\text{Th})$ can be determined by the y-intercept of the $(^{238}\text{U}/^{232}\text{Th})_{\text{M}}$ vs. $(^{230}\text{Th}/^{232}\text{Th})_{\text{M}}$ Rosholt isochron or the slope of the $(^{232}\text{Th}/^{238}\text{U})_{\text{M}}$ vs. $(^{230}\text{Th}/^{238}\text{U})_{\text{M}}$ Osmond isochrons (Lin et al., 1996; Blard et al., 2011). Various methods have been investigated to construct sample pair isochrons. Initially studies relied on leachate/residue (L/R) and leachate/leachate (L/L) methods (e.g., Kaufman, 1971; Schwarcz and Latham, 1989) to produce measurement pairs used to construct isochrons. However, leachate schemes have been found to preferentially solubilize U or Th. If the majority of U is in the authigenic carbonate phase these methods may be acceptable (e.g., Ku and Liang, 1984; Schwarcz and Latham, 1989; Przybłowicz et al., 1991; Kaufman, 1993). Alternatively, to produce consistent mixing relationships between the authigenic and detrital components a Total Sample Dissolution (TSD) method, involving full digestion of suites ($n \geq 3$) of coeval

samples (Bischoff and Fitzpatrick, 1991; Luo and Ku, 1991), has been used in recent studies (e.g., Hall and Henderson, 2001; Soligo et al., 2002; Garnett et al., 2004; Haase-Schramm et al., 2004; Blard et al., 2011; Torfstein et al., 2013). In this study we use the TSD method.

Several methodologies have been proposed to construct statistically significant linear regressions of the paired isochrons. Statistical rigor is required to produce meaningful and representative ^{230}Th -U ages while accounting for both analytical uncertainties in MC-ICPMS or TIMS measurements, as well as geologic scatter. Luo and Ku (1991) proposed a methodology using simple least-squares fit with counting statistics which does not consider error-weighting or error correlation (in x and y errors), noting that doing so would require more than three or four coeval samples, often impractical given sample size and laboratory method limitations (cf. Ku, 2000). Ludwig and Titterton (1994) propose a 3-D simultaneous solution of the Osmond isochrons using maximum-likelihood estimation (MLE), allowing for the projection of the detritus free end-member onto a traditional ^{230}Th - ^{234}U - ^{238}U evolution plot. This method is implemented in ISOPLOT (Ludwig et al., 2012). Finally, recent studies (e.g., Hall and Henderson, 2001; Soligo et al., 2002; Garnett et al., 2004; Blard et al., 2011) have used error-weighted 2-D linear fits to the individual Rosholt or Osmond isochrons using methods originally developed by York (1969) and implemented in ISOPLOT (Ludwig et al., 2012).

For the suites of coeval samples, to calculate $(^{230}\text{Th}/^{238}\text{U})_A$ and $(^{234}\text{U}/^{238}\text{U})_A$, and calculate a detrital Th corrected isochron age, we evaluated both solutions to separate 2-D isochrons (e.g., Hall and Henderson, 2001; Soligo et al., 2002; Garnett et al., 2004; Blard et al., 2011) and the simultaneous 3-D solution of the Osmond isochrons using maximum-likelihood estimation (MLE) (Ludwig and Titterton, 1994). Both methods calculate nearly identical absolute ages, but differ in the calculated error on $(^{230}\text{Th}/^{238}\text{U})_A$ and $(^{234}\text{U}/^{238}\text{U})_A$. For the 2-D solutions, to best illustrate spread along the isochrons we choose to use the slopes of Rosholt isochrons ($(^{238}\text{U}/^{232}\text{Th})_M$ vs. $(^{230}\text{Th}/^{232}\text{Th})_M$ and $(^{238}\text{U}/^{232}\text{Th})_M$ vs. $(^{234}\text{U}/^{232}\text{Th})_M$). While constructing isochrons some sub-samples were rejected to minimize the error on the slope (the calculated $(^{230}\text{Th}/^{238}\text{U})_A$ used in equation (1)) of the $(^{238}\text{U}/^{232}\text{Th})_M$ vs. $(^{230}\text{Th}/^{232}\text{Th})_M$ Rosholt isochron and produce a mean square

weighted deviation (MSWD) closest to ~ 1 (cf. Hall and Henderson, 2001; Garnett et al., 2004). In an effort to maximize the precision of our ^{230}Th -U ages we attempted to construct isochrons using 5 to 7 sub-samples, more than previous studies that have used isochron approaches to date impure carbonates and evaporites (e.g Luo and Ku, 1991; Blard et al., 2011).

3.3.4. Radiocarbon analyses

Tufa material was sampled for radiocarbon dating directly adjacent to 15 samples processed for coeval U-series analysis. By using an optical microscope, porous textures and visible organic/siliciclastic material were avoided. Samples underwent a mild acid etch, prior to conversion to graphite and analysis by accelerator mass spectrometry (AMS) at Beta Analytic Inc. in Miami, Florida. Analyses were adjusted for $\delta^{13}\text{C}$ content. To directly compare the radiocarbon results with U-series results all ages are also reported as calibrated ages BP (2σ) (age range, median age and mean age) using the extended IntCal09 calibration curve processed using the program Calib 6.0 (Stuiver and Reimer, 1993; Reimer et al., 2009).

3.4. Climate model output analysis

To calculate changes in precipitation and evapotranspiration predicted by AOGCMs for the western United States, we computed average monthly climatologies using the monthly precipitation, total evaporation, runoff, relative humidity and mean annual temperature outputs for the LGM (21 ka boundary conditions) and pre-industrial control experiments available via the Paleoclimate Model Intercomparison Project 3 (PMIP3) (Braconnot et al., 2012; <http://pmip3.lsce.ipsl.fr/>). The nine analyzed PMIP3 models include NCAR-CCSM4 (Gent et al., 2011; Brady et al., 2013), CNRM-CM5 (Voldoire et al., 2013), FGOALS-g2, IPSL-CM5A-LR (Kageyama et al., 2013a,b), MRI-CGCM3 (Yukimoto et al., 2012), MPI-ESM-P, GISS-E2-R, COSMOS-ASO and MIROC-ESM. These models were used because of their inclusion into the PMIP3/CMIP5 database (Coupled Model Intercomparison Project 5; <http://cmip-pcmdi.llnl.gov/cmip5/>). Monthly and annually summed precipitation, total evaporation and runoff anomalies (LGM minus pre-industrial control) were calculated for Surprise Valley (120°W , 41.5°N) using bilinear interpolation for both climate model experiments.

Similarly, annually averaged relative humidity and temperature anomalies were calculated. Not all output variables are uniformly archived in the PMIP3 database. COSMOS-ASO does not report runoff values, and COSMOS-ASO, FGOALS-g2, MIROC-ESM and MPI-ESM-P do not report relative humidity values. No bias correction was applied to the climate models outputs; instead, for comparison to paleoclimate records we report the percent change simulated by the LGM experiments minus pre-industrial control experiments for each climate model in the PMIP3 ensemble.

4. RESULTS

In the following sections, we present and examine the reliability of the geochronologic approaches, spatial relationships, and the U-series and stable isotope systematics. We then construct the late Pleistocene hydrograph for Lake Surprise. For the purposes of discussion, the tufa samples are split into three distinct groups separated by elevation (and age): (1) samples collected from the prominent lower elevation shorelines at 1419 to 1472 m; (2) samples collected from the less-prominent middle elevation shorelines of 1478 to 1531 m; and (3) samples collected from higher elevation shorelines >1542 m (**Figure 2B-E**).

4.1. Evaluating stable isotope, Sr/Ca and U-series systematics

4.1.1. Stable isotopic and elemental signatures of tufas and modern samples

Measurements of $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and Sr/Ca in Lake Surprise carbonates and modern Surprise Valley waters and sediments provides a framework for evaluating the validity and spatial relationship of the ages obtained in this study. All samples from <1531 m have $\delta^{18}\text{O}$ values (VPDB) of -2.50 to -4.51 ‰ (average $\delta^{18}\text{O}$ = -3.46 ± 0.78 ‰ (2σ); $n=32$), and $\delta^{13}\text{C}$ values of 2.68 to 4.10 ‰ (average $\delta^{13}\text{C}$ = 3.59 ± 0.49 ‰ (2σ); $n=32$) (modern playa carbonate: $\delta^{18}\text{O}$ = -4.05 ‰ and $\delta^{13}\text{C}$ = -0.11 ‰; $n=2$). High elevation samples have much lower $\delta^{18}\text{O}$ values of -13.13 to -9.33 ‰ ($n=11$), as well as lower $\delta^{13}\text{C}$ values of -6.32 to -1.72 ‰ ($n=10$). Ingraham and Taylor (1989) report a large, -14.4 ‰ seasonal variation in modern lake water $\delta^{18}\text{O}$ from the middle playa (**Figure 2A**) of -9.5 to 4.9 ‰ (VSMOW) with minimal variation in creek waters, from -15.0 to -13.7 ‰ (Vienna Standard Mean Ocean Water, VSMOW) (Ingraham and Taylor, 1989; Sladek et al., 2004) (**Table A5**).

Assuming the range of measured modern water Sr/Ca (**Table 4**) reflects paleolake water, and a range of potential partition coefficients of $K_d^{Sr} = \frac{(Sr/Ca)^{calcite}}{(Sr/Ca)^{solution}} = 0.12 - 0.35$ (Gabitov and Watson, 2006), the Sr/Ca measured in the tufa samples (**Table 3**) is in the higher range of expected values. The group 3 samples have a lower Sr/Ca (average Sr/Ca = 0.86 ± 0.66 mmol/mol (2σ); n=7) compared to the samples from elevations <1531 m (average Sr/Ca = 1.16 ± 0.30 mmol/mol (2σ); n=26).

A positive covariance with a rating of weak, moderate or strong is reported in **Table 3**; the rating follows the method of Davis et al. (2009) and uses a Pearson product-moment coefficient (r^2) in $\delta^{18}O$ - $\delta^{13}C$ and $\delta^{18}O$ -Sr/Ca. For all samples <1531 m (n=29) both $\delta^{18}O$ - $\delta^{13}C$ ($r^2 = 0.47$) and $\delta^{18}O$ -Sr/Ca ($r^2 = 0.36$) are moderately correlated (**Figure 3C and 3D**). For group 1 samples of LGM age (n=19), $\delta^{18}O$ - $\delta^{13}C$ is strongly correlated ($r^2 = 0.61$); however, $\delta^{18}O$ -Sr/Ca ($r^2 = 0.06$) is not correlated, likely due to minimal spread in measured Sr/Ca ratios (**Figure 3A and 3B**). For group 2 samples (n=7) $\delta^{18}O$ - $\delta^{13}C$ ($r^2 = 0.40$) is moderately correlated and $\delta^{18}O$ -Sr/Ca ($r^2 = 0.92$) is very strongly correlated (**Figure 3A and 3B**). Assessment of $\delta^{18}O$, $\delta^{13}C$ and Sr/Ca from the group 3 (highest elevation) outlier samples suggests potential alteration of the samples due to recrystallization or pedogenic overprinting (**Table 3, Figure 3**).

In addition to the modern topographic controls (**Figure 2A, Table 2**), the covariance of $\delta^{18}O$ - $\delta^{13}C$ and $\delta^{18}O$ -Sr/Ca can also be used to evaluate the role of evaporation in lake systems (Müller et al., 1972; Eugster and Kelts, 1983; Talbot, 1990; Li and Ku, 1997; Davis et al., 2008, 2009; Horton and Oze, 2012; Chamberlain et al., 2013). Terminal basin lakes have comparatively longer residence times, resulting in the evaporative enrichment of ^{18}O in the lake water and preferential outgassing of ^{12}C -rich CO_2 from the system (Talbot, 1990). Evaporative effects on $\delta^{18}O$ and $\delta^{13}C$ require that if a lake is hydrologically closed and evaporation is high, then the $\delta^{18}O$ and $\delta^{13}C$ will covary (Talbot, 1990; Li and Ku, 1997; Davis et al., 2009; Horton and Oze, 2012). Similarly, Sr/Ca ratios have also been used to assess evaporative effects in lake systems (Muller et al., 1972; Davis et al., 2009). Muller et al. (1972) show that for highly evaporative lakes, Sr/Ca in carbonates is proportionally correlated to the Sr and Ca concentrations in lake water. With a partition coefficient (K_d) less than 1 (Gabitov and

Watson, 2006), Sr is not removed as efficiently from lakewater as Ca, which is taken up by precipitating carbonates; thus, Sr is concentrated when the net evaporative flux, relative to lake volume, is high (Eugster and Kelts, 1983; Davis et al., 2009). This correlation is not expected to hold with an open system lake because Sr would be flushed from the system. Surprise Valley tufa samples demonstrate a moderate to high $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ -Sr/Ca covariance among sample groups (**Table 3, Figure 3**). While application of the covariance of $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ -Sr/Ca in shoreline tufa has not been extensively explored, based on previous studies of lacustrine carbonates (e.g., Talbot and Kelts, 1990; Li and Ku, 1997; Davis et al., 2009; Horton and Oze, 2012), and the relatively robust trends observed here (**Table 3, Figure 3**), we propose that the $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ -Sr/Ca system is responding to the closed basin nature of Lake Surprise.

The extent to which the Surprise Valley lake system was a terminal basin in the past, with no outflow or substantial regional groundwater input, is important for using lake shoreline ages (and other geochemical data) as indicators of past climate. Coupling light stable isotopic and elemental analysis with modern watershed data demonstrates that, indeed, the latest Pleistocene Lake Surprise lake system was a closed, inward-draining pluvial lake system.

4.1.2. U-series measurements of modern samples

The 19 water samples and 5 modern playa samples (see **Figure 2A** for locations) demonstrate the range of expected variability in U concentrations and ($^{234}\text{U}/^{238}\text{U}$) within the Surprise Valley watershed (**Table 4**). The average measured U concentrations in surface waters and groundwaters from Surprise Valley are 0.430 ± 1.722 ng/ml (2σ) ($n=8$) and 0.687 ± 1.722 ng/ml (2σ) ($n=5$), respectively. Hot spring waters have much lower U concentrations, ranging from 0.0031 to 0.2552 ng/ml (average [^{238}U] = 0.0770 ± 0.197 ng/ml (2σ); $n=6$), suggesting that hot springs contribute minimally to the U budget of Lake Surprise. Surface water ($^{234}\text{U}/^{238}\text{U}$) measurements, draining from primarily basaltic bedrock, range from 1.478 to 2.360 (average ($^{234}\text{U}/^{238}\text{U}$) = 2.076; $n=8$). The ($^{234}\text{U}/^{238}\text{U}$) from groundwater and hot spring waters demonstrate a greater range of variability from 1.391 to 3.508 (average ($^{234}\text{U}/^{238}\text{U}$) = 2.187; $n=7$). Modern playa pore waters (($^{234}\text{U}/^{238}\text{U}$) = 1.455 - 1.640; $n=4$), authigenic carbonate fractions (($^{234}\text{U}/^{238}\text{U}$) = 1.327 - 1.571; $n=4$) and the modern playa carbonate (($^{234}\text{U}/^{238}\text{U}$) = 1.441; $n=1$) fall within

the lower end of the range of observed water sample variability.

4.2. Geochronology

4.2.1. ^{230}Th -U geochronology

The results of the U and Th isotope measurements, and the ^{230}Th -U age calculations are presented in **Tables 5** and **6** and **Table A3**. **Figure 4** and **Figures A3-5** demonstrate the ^{230}Th -U age calculations via pairs of 2-D Rosholt isochrons and 3-D Osmond isochrons for suites of coeval samples. For all suites of 5+ coeval samples, calculated Rosholt isochron ages are reported in **Table 5**. One important assumption is that these isochrons constitute mixing relationships between the authigenic carbonate end-member and a Th end-member with a distinct Th isotope composition, either detrital or hydrogenous Th. Hydrogenous ^{232}Th in modern surface waters was less than 6 pg/ml, which suggests that detrital Th is the only substantial Th isotope end-member for Lake Surprise carbonates, and is likely introduced into the tufa via incorporation of siliciclastic or organic particles containing Th.

To calculate the most robust isochron fit that minimizes the error on the calculated $(^{230}\text{Th}/^{238}\text{U})_A$ and the eventual ^{230}Th -U age, some sub-samples were excluded from the error-weighted linear regressions (these sub-samples are indicated in **Table A3**). For three samples (SVDI11-T2 Outer Rind, SVDI11-T18 and SVDI12-T12) no robust 2-D Rosholt or 3-D Osmond isochrons could be calculated. Additionally, two samples (SVDI11-T14 and SVDI12-T9) produced no robust 3-D Osmond isochron solutions because the calculated ^{230}Th -U age error was greater than the corrected ^{230}Th -U age. After excluding outlying data points, all Rosholt and Osmond ages are concordant with nearly identical calculated ^{230}Th -U ages (**Figure 5A**). Differences in the age error between the two methods are the result of how analytical uncertainty and geologic scatter are handled (cf. Ludwig and Titterton, 1994, Ku, 2000). To be consistent with ^{230}Th -U isochron ages reported in recent literature (e.g., Hall and Henderson, 2001; Soligo et al., 2002; Garnett et al., 2004; Blard et al., 2011), hereafter we will refer to the ^{230}Th -U ages from the Rosholt isochrons.

For all 111 sub-samples we calculated single sample (SS) detrital Th corrected ages. For samples with robust 2-D Rosholt isochrons (n=68), we use the y-intercept of

the $(^{238}\text{U}/^{232}\text{Th})_{\text{M}}$ vs. $(^{230}\text{Th}/^{232}\text{Th})_{\text{M}}$ to calculate the $(^{230}\text{Th}/^{232}\text{Th})_0$ used in equation (3) (e.g., Torstein et al., 2013). For the remaining samples (n=43) an error-weighted mean of the isochron intercepts and the modern carbonate $(^{230}\text{Th}/^{232}\text{Th})$ measurement was used to calculate a basin average $(^{230}\text{Th}/^{232}\text{Th})_0$ ($(^{230}\text{Th}/^{232}\text{Th})_0 = 1.327 \pm 0.096$, error-weighted mean, **Table A4**). The calculated value is similar to other Great Basin paleolakes (Lin et al., 1996; Lin et al., 1998). All age calculations and the error-weighted means of the SS detrital Th corrected ages for suites of sub-samples (n=2 to 7) were calculated using ISOPLOT 3.75 (Ludwig, 2012).

Comparing the error-weighted means of the SS detrital Th corrected ^{230}Th -U ages to the Rosholt isochron ^{230}Th -U ages (consisting of only samples from groups 1 and 2), indicates the utility of the two complementary age calculation methods. For samples with robust isochrons, all samples are concordant (**Figure 5A**) or nearly concordant. On average, the error-weighted mean SS ages are 0.17 kyrs younger than the Rosholt ages, with no apparent trend with elevation (**Figure 5B**). The largest outlier is SVDI12-T1, which has a much older Rosholt isochron age (26.44 ± 5.86 ka BP compared to 18.54 ± 0.88 ka BP). The older age is a result of a low $(^{230}\text{Th}/^{232}\text{Th})_0$ isochron intercept and steeper slope (yielding an older age) in the $(^{238}\text{U}/^{232}\text{Th})_{\text{M}}$ vs. $(^{230}\text{Th}/^{232}\text{Th})_{\text{M}}$ isochron. Due to this anomalously low $(^{230}\text{Th}/^{232}\text{Th})_0$ the basin average $(^{230}\text{Th}/^{232}\text{Th})_0$ was used to calculate the SS ages for SVDI12-T1. SVDI12-T2, from the same elevation (1419.5 m), yielded nearly concordant ages of 22.78 ± 1.23 ka BP (Rosholt age) and 24.84 ± 0.92 ka BP (SS age).

The calculated $(^{234}\text{U}/^{238}\text{U})_0$ of the tufa samples (see **Tables 5 and A3**), which should reflect the $(^{234}\text{U}/^{238}\text{U})$ of the water precipitating carbonate, demonstrate less variation in $(^{234}\text{U}/^{238}\text{U})$ than modern water, carbonate and playa samples, and form three distinct ranges of $(^{234}\text{U}/^{238}\text{U})_0$ (**Figure 6**). The three observed groups are: (1) the lower elevation (1420-1478 m) samples of ~18 to 27 ka, clustering around a $(^{234}\text{U}/^{238}\text{U})_0$ range of 1.497 to 1.682; (2) the middle elevation (1509-1531 m), post-LGM samples with elevated $(^{234}\text{U}/^{238}\text{U})_0$ from 1.784 to 1.846; and (3) the highest elevation samples (>1542 m), with $(^{234}\text{U}/^{238}\text{U})_0$ between 1.128 to 1.261, below the minimum of modern values (**Figures 2B-E and 6**).

4.2.2. Radiocarbon geochronology

The radiocarbon geochronology results and Calib 6.0 (Stuvier and Reimer, 1993) calibrations are presented in **Table 6**. Seven of the fifteen samples have ^{14}C ages that intersect the IntCal09 calibration curve (Reimer et al., 2009) at multiple ages (2σ significance), resulting in two or more statistically probable calibrated ages. This is primarily the result of plateaus and variations in atmospheric ^{14}C production documented in the IntCal09 calibration during the last deglaciation (see **Figures A6-9**). For all samples the ages discussed hereafter are the calibrated average ages of the tufa samples taken from the age range with the highest relative probability based on the area under the calibrated age probability distribution. For all but two samples (SVDI11-T4-1b and SVDI12-T15-b) this area was $>88\%$ of the total area under the curve for the age (**Table 6**).

Plotting radiocarbon age against elevation, the construction of a hydrograph using only the radiocarbon ages (**Figure 7B**) indicates that twelve of the fifteen radiocarbon ages record the transgression and regression of Lake Surprise. Two samples, SVDI12-T1-a and SVDI12-T2-b, from the lowest shoreline (1419.5 m) of the Middle Lake shoreline set are not concordant but have very similar ages of 20.91 ± 0.40 and 21.82 ± 0.34 ka cal BP, respectively. Similarly, two nearly identical elevation samples, SVDI11-T3-2 (1437.7 m) and SVDI12-T13 (1437.2 m), from different sampling localities have concordant ages of 20.92 ± 0.41 and 20.85 ± 0.42 ka cal BP. Lower elevation samples (sample group 1), with the exception of one anomalously younger age (SVDI12-T5-b), record transgression from 1419.5 to 1453.5 m, and have LGM ages ranging from 19.13 ± 0.26 to 21.82 ± 0.34 ka cal BP. Four tufa samples at middle elevations (1530.7 to 1478.4 m, sample group 2) and the anomalous low elevation sample (1444.3 m), record consistent gradual regression from an apparent highstand at 15.23 ± 0.36 to 10.69 ± 0.11 ka cal BP. Outliers to this apparent lake cycle are three group 3 samples (SVDI11-T18-1c, SVDI12-T12-1 and SVDI12-T18) from three of the highest shoreline elevations (1556, 1577 and 1564 m respectively), at different sample localities, which have discordant ages of 8.56 ± 0.07 , 16.33 ± 0.47 and 25.00 ± 0.45 ka cal BP. Of these three samples, the lowest elevation sample (SVDI12-T18 at 1556 m), recorded an age of 16.33 ± 0.47 ka cal BP, which is potentially compatible with lower elevation samples.

4.2.3. Comparison of radiocarbon and ^{230}Th -U geochronology

Studies attempting to reconcile radiocarbon ages and ^{230}Th -U ages in lacustrine carbonates, particularly tufa samples, have considered the limitations of both methods (e.g., Lin et al., 1996; Placzek et al., 2006a; Kurth et al., 2011; Zimmerman et al., 2012). Radiocarbon dates can be influenced by young carbonate contamination, resulting in an anomalously young age (Cassata et al., 2010), or conversely, dead carbon incorporation, resulting in an anomalously old age (Zimmerman et al., 2012). The Surprise Valley watershed, unlike other watersheds containing late Pleistocene lakes, drains primarily basaltic bedrock and contains no substantial carbonate sediments (Egger and Miller, 2011). Thus, radiocarbon dates are unlikely to be affected by dead carbon incorporation. For the samples dated in this study, three to five times the ^{14}C half-life ($t_{1/2} = 5.73$ kyrs), the offset of radiocarbon ages caused by modern contamination up to 5% can be significant to as much as 7 kyr (cf. Cassata et al., 2010, Figure 4). The potential for open-system behavior in the uranium system, typically a result of U-loss due to the preferential solubility of U relative to Th, results in an anomalously high ($^{230}\text{Th}/^{238}\text{U}$) and an older apparent age (cf. Kurth et al., 2011). Given these constraints, for samples dated by both methods, we consider the radiocarbon ages (**Table 6**) as minimum ages and the ^{230}Th -U ages (**Table 5**) as maximum ages.

Eleven samples dated using both radiocarbon and ^{230}Th -U demonstrate acceptable agreement. Seven of the eleven sample ages (Rosholt ^{230}Th -U ages) are concordant between the two methods, and the remaining four have ^{230}Th -U ages that are on average 1.35 kyr older (**Figure 5C and 5D**). Previous studies on other paleolakes have also observed that a majority of samples dated using both methods, while falling close to the 1:1 concordant line, have slightly older ^{230}Th -U ages (e.g., Kaufman and Broecker, 1965; Placzek et al., 2006a; Blard et al., 2011).

The only significant outlier from the lower elevation samples is SVDI12-T5. This tufa sample, formed on highly vesicular basalt bedrock, has a significantly younger radiocarbon age of 10.69 ± 0.11 ka cal BP, compared to concordant ^{230}Th -U ages of 22.87 ± 4.17 ka BP (Rosholt age) and 24.91 ± 3.10 ka BP (SS age). This suggests either an insufficient detrital Th correction or modern contamination of the carbonate sample. The only concordant group 3 sample age is the SS average of SVDI11-T18 with a

radiocarbon age of 16.33 ± 0.47 ka cal BP and a ^{230}Th -U age of 15.41 ± 5.85 ka BP; however, the SS ages range from 6.30 to 20.74 ka BP.

Taking the eleven accepted radiocarbon ages as the ‘true’ age of the samples also allows for the comparison between the ages calculated using isochron or single sample methods (**Figure 5B**). SS ^{230}Th -U ages are on average 0.98 kyr older than the radiocarbon ages, and are closer than the Rosholt ^{230}Th -U isochron ages. The SS detrital correction method only corrects the ($^{230}\text{Th}/^{238}\text{U}$) ratio (via equation (3)), whereas Rosholt ^{230}Th -U isochrons also correct the ($^{234}\text{U}/^{238}\text{U}$) of the authigenic carbonate (cf. Ku, 2000). Although the correction of the ($^{234}\text{U}/^{238}\text{U}$) of the authigenic carbonate is negligible at the young age range of these samples, we consider the Rosholt ^{230}Th -U isochron ages as the most accurate ^{230}Th -U ages of the samples, because this method corrects both the U and Th ratios of the authigenic carbonate for detritus. Additionally, depending on the ($^{232}\text{Th}/^{238}\text{U}$) ratio of a given sub-sample, the age error of the SS error-weighted ^{230}Th -U ages will vary significantly. However, given the concordance of Rosholt isochron and error-weighted averages of the SS ^{230}Th -U ages (**Figure 5B**) we consider the SS ^{230}Th -U ages as supporting ages in the hydrograph construction.

4.2.4. Older pre-LGM ages from high shorelines

Two samples from the highest shoreline elevations, and the Upper Lake shoreline set, record ages that may be from past pluvial lake cycles; however, both samples recorded ($^{234}\text{U}/^{238}\text{U}$)₀ less than the range of modern water and playa samples. An SS age from a group 3 sample 1542.3 m yielded an early MIS 4 age of 70.37 ± 1.55 ka BP. Additionally, one sub-sample of SVDI12-T18b at 1564.2 m (**Table 1; Figure 2B-E**) recorded an early MIS 5 age of 115.92 ± 6.44 ka BP (**Table A3**). This age is likely questionable, as the other sub-sample appears to have suffered from open system effects, recording potential U-loss, evidenced by a much lower U concentration (417 vs. 2544 ppb) and an anomalously high ($^{230}\text{Th}/^{238}\text{U}$) of 2.277, resulting in an incalculable age. This open system behavior is also reflected in a discordant radiocarbon age of 25.00 ± 0.45 ka cal BP. While inconclusive, these two samples may record a higher, older lake cycle(s) in Surprise Valley, similar to those found in other Basin and Range lake systems, corresponding to MIS 4 or MIS 6 (Reheis, 1999; Kurth et al., 2011). Corroborating these conclusions is evidence from modern fish species, suggesting that the species found in

the western tributaries and alkali lakes of Surprise Valley are closely related to fish found in both the Warner and Lahontan systems, implying that the lakes were linked pre-MIS 2 (Hubbs and Miller, 1948; Moyle and Daniels, 1982).

Figure 3. Correlation plots of Sr/Ca- $\delta^{18}\text{O}$ (**A** and **C**) and $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ (**B** and **D**) for tufa samples grouped by elevation. Sample are grouped using the same symbology as in **Figure 2B-E**. Pearson product-moment coefficients (r^2) were calculated for linear least squares regressions for Sr/Ca- $\delta^{18}\text{O}$ and $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ for each sample group and for the combination of sample groups 1 and 2. Combining sample groups 1 and 2 (**A-B**) demonstrates similar Sr/Ca- $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ systematics among all samples <1531 m Group 3 samples have very different $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ signatures (**C-D**), suggesting potential recrystallization and/or different source water origins.

Figure 4. Example of Rosholt and Osmond isochron plots calculating ^{230}Th -U ages for SVDI12-T2. All error ellipses and error envelopes are 2σ . **(A-B)** Rosholt isochrons of $(^{230}\text{Th}/^{232}\text{Th})$ vs. $(^{238}\text{U}/^{232}\text{Th})$ and $(^{234}\text{U}/^{232}\text{Th})$ vs. $(^{238}\text{U}/^{232}\text{Th})$ (measured values), with error-weighted linear regressions (see text for details), calculates the $(^{230}\text{Th}/^{238}\text{U})_{\text{authigenic}}$ and $(^{234}\text{U}/^{238}\text{U})_{\text{authigenic}}$, based on the slopes of the linear regressions. The intercept of $(^{230}\text{Th}/^{232}\text{Th})$ vs. $(^{238}\text{U}/^{232}\text{Th})$ is the $(^{230}\text{Th}/^{232}\text{Th})_{\text{initial}}$ of the suite of coeval samples. Samples not included in linear regressions are gray ellipses. **(C)** Projected representation of a 3-D Osmond isochron calculation of ^{230}Th -U ages for coeval suites of samples (see text for details) onto a ^{230}Th - ^{234}U - ^{238}U evolution plot. Black ellipses are measured values and white ellipses are projected values based on the projected isochron (gray line). Gray ellipses are the calculated ^{230}Th -U age (95% confidence) and $(^{234}\text{U}/^{238}\text{U})_0$ of the sample.

Figure 5. Comparison of paired calibrated radiocarbon (^{14}C) ages to $^{230}\text{Th-U}$ ages and comparison among $^{230}\text{Th-U}$ age calculations (see text for details). All error ellipses are 2σ . **(A)** Rosholt Isochron based $^{230}\text{Th-U}$ ages compared to 3-D Osmond isochron based $^{230}\text{Th-U}$ ages. **(B)** Rosholt Isochron based $^{230}\text{Th-U}$ ages compared to error-weighted averages of single sample (SS) detrital corrected $^{230}\text{Th-U}$ ages. **(C)** Rosholt Isochron based $^{230}\text{Th-U}$ ages to radiocarbon (^{14}C) ages. **(D)** Error-weighted averages of single sample (SS) detrital corrected $^{230}\text{Th-U}$ ages compared to radiocarbon (^{14}C) ages.

Figure 6. Comparison of ^{230}Th -U age equation (equation (2)) calculated $(^{234}\text{U}/^{238}\text{U})_0$ to measured modern $(^{234}\text{U}/^{238}\text{U})$. Only sample groups 1 and 2 are plotted. **(A)** Rosholt isochron derived $(^{234}\text{U}/^{238}\text{U})_0$ are shown as ellipses with the range values for the single sample detrital Th corrected samples shaded in gray. All errors are 2σ . **(B)** Box and whisker plots show the median and interquartile ranges of modern $(^{234}\text{U}/^{238}\text{U})$ and the rejected single sample ages' $(^{234}\text{U}/^{238}\text{U})_0$ from the group 3 samples.

4.3 Lake Surprise hydrograph

To construct the Lake Surprise hydrograph we use only ages obtained from sample groups 1 and 2 and plot calibrated radiocarbon ages (**Table 6**), and both SS ^{230}Th -U ages and Rosholt isochron ^{230}Th -U ages (**Table 5**) against the lidar-derived sample elevations (**Table 1**). We connected radiocarbon ages and Rosholt isochron ^{230}Th -U ages separately, assuming that a given sample records the minimum lake water level (e.g., Felton et al., 2006; Benson et al., 1996; 2011; Zimmerman et al., 2012). For samples of statistically equivalent ages, the sample of higher elevation was taken as the likely lake

level, rather than recording higher frequency fluctuations. In addition to the radiocarbon and ^{230}Th -U based hydrographs, we constructed an age envelope of all available ages from samples less than 1531 m (**Figure 7B**). Based on the radiogenic and stable isotopic evidence presented here, we have concluded that Lake Surprise lake levels reached 1531 m (a water depth of 176 m) at 15.23 ± 0.36 ka cal BP (^{14}C age) during the latest Pleistocene (**Figure 7B**). This highstand age is supported by a concordant, but significantly older, ^{230}Th -U isochron age of 18.25 ± 2.67 ka BP (Rosholt age). At higher elevations, one sample (SVDI11-T18), suggests that Lake Surprise may have briefly reached 1555 m at 16.33 ± 0.47 ka cal BP (^{14}C age), although previously presented evidence suggests that this age may be in error due to calcite recrystallization causing open system behavior.

Our results correlate well with the limited previous studies, and help clarify some ambiguous relationship (**Table A1**). The infrared-stimulated luminescence (IRSL) age on feldspar grains constraining the older age range of the deltaic deposits from Personius et al. (2009) appears to be too old (28.08 ± 4.44 ka), as originally suspected by the authors, due to potential contamination. However the younger age of 18.40 ± 3.17 ka of the deltaic sediments, using a weighted mean of the optically stimulated luminescence (OSL) and IRSL ages, (Personius et al., 2009), agrees well with the constructed lake level curve. Finally, three of the four subaerial ages from Personius et al. (2009) are much younger than the proposed lake level at that elevation (1475 m); however, the oldest sub-aerial age, 13.38 ± 2.64 ka is within error of the plotted radiocarbon and ^{230}Th -U based hydrographs during the deglacial lake level recession. All other previous constraints on Lake Surprise lake level are consistent with the constructed hydrograph plotted in **Figure 7B**.

Figure 7. (A) Calculated precipitation increase (% modern) from the isotope mass balance model. See **Tables 7** and **8** for assumptions and calculations. (B) New hydrograph for the latest pluvial cycle of Lake Surprise. The gray error envelope bounds all sample ages with elevations <1531 m, the gray line relies on radiocarbon ages only and black line relies on Rosholt ²³⁰Th-U ages. Dashed lines are tentative correlations. Other constraints include moderate to deep water deposition of the Trego Hot Spring tephra at 1378m (Hedel, 1980; 1984), Cooks Canyon deltaic lacustrine and subaerial sediments deposited at 1493m (Personius et al., 2009) and playa prehistoric settlements (O'Connell and Inoway, 1994). See **Figure 2A** for the location of the additional lake level constraints.

5. DISCUSSION

In the following discussion, we examine the Lake Surprise hydrograph of **Figure 7B** within the context of previous work and explore further the implications for western United States paleoclimatology during the late Pleistocene. The key results include: (1) Lake Surprise's watershed remained a closed, inward draining basin throughout the last glacial cycle and lateral correspondence was found among samples from the four shoreline localities based on topographic analysis, $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ -Sr/Ca covariance and concordant geochronology (radiocarbon and U-series); (2) the new lake hydrograph (**Figure 7B**), in agreement with previous lake level constraints (**Table A1**), places the highest lake level ~176 m above modern playa at >15.23 ka and during the Last Glacial Maximum at moderate levels (~80 m above modern playa); and (3) multiple lines of evidence reveal that samples from the highest shorelines are likely from older, higher lake cycles and were influenced by variable amounts of open-system exchange or pedogenic overprinting.

Expanding upon these results, we modify an existing $\delta^{18}\text{O}$ isotope mass balance model by including basin geometry to assess hydrologic controls on Lake Surprise lake levels and compare the calculated changes in precipitation (**Figure 7A**) to climate models from the PMIP3 climate model ensemble (**Figure 8**). Finally, we discuss the implications of the latest lake cycle of Lake Surprise and the results from the isotope and climate modeling in the context of regional and global climate during the last deglaciation.

5.1 Hydrologic controls on Lake Surprise lake levels

To constrain changes in atmospheric circulation patterns and the seasonal role of solar insolation on regional climate during the last deglaciation requires quantitative predictions of meteorological and hydrologic processes. To constrain these processes, we incorporate the Lake Surprise hydrograph, the stable isotope analyses and the topographic analysis of the basin geometry into two models that enable us to estimate past changes in precipitation and evaporation.

5.1.1. Hydrologic index calculations

The “Pluvial Hydrologic Index” (*HI*), originally derived by Mifflin and Wheat (1979), is a measure of the ratio of lake surface area to tributary area given an equilibrium (steady-state) lake level, whereby moisture into the lake equals moisture out

of the lake. It has been used in various forms (e.g., Hostetler and Benson, 1990; Reheis, 1999b) to solve, to first order, for the ratio of runoff to net lake evaporation using paleohydrologic constraints derived from related landforms (e.g., wave-cut shorelines, beaches, strandlines) preserved in the geomorphic record. Based only on mass balance, the dimensionless hydrologic index (*HI*) is (Mifflin and Wheat (1979)):

$$HI = \frac{A_L}{A_T} = \frac{A_L}{A_B - A_L} = \frac{R_T}{E_L - P_L} = \frac{P_T - ET_T}{E_L - P_L} \quad (4)$$

where A_L , A_B and A_T are the areas of the lake, basin and tributary ($A_T = A_B - A_L$), P is the average on-lake (subscript L) and tributary (subscript T) precipitation, E_L is the gross lake evaporation, ET_T is the average tributary evapotranspiration and R_T is the combined surface and groundwater runoff, where $R_T = P_T - ET_T$. Using the modern topographic analysis of Surprise Valley we have computed the *HI* from A_L/A_T at (1) the range of LGM stillstands (1420-1440 m), (2) the post-LGM highstand (1531 m), and (3) the pre-MIS 2 (possibly MIS 6) highstand (1567 m, documented by Mifflin and Wheat (1979) and Reheis (1999), and confirmed these values using the lidar dataset in the accommodation zone shoreline set, **Figure 2B**) (**Table 2**). Modern meteorology ($(P_T - ET_T)/(E_L - P_L)$) and modern lake levels (A_L/A_T) indicate that the *HI* equals 0.13 and 0.12, respectively. During the LGM stillstands, the *HI* was almost 300% of modern, with values of 0.29 to 0.34. At the post-LGM deglacial highstand, the *HI* was 0.56. At its maximum, the pre-MIS 2 highstands indicate an *HI* of ~0.65. The calculated *HI* values are consistent with other nearby lake systems evaluated by Mifflin and Wheat (1979) in south-central Oregon and northwestern Nevada and provide a useful framework to constrain changes in precipitation, evaporation and temperature during the LGM (e.g., Reheis, 1999b).

5.1.2. Isotope mass balance calculations

Stable isotopes provide a complementary method to the *HI* calculations for determining basin average changes in precipitation. We modify the isotope mass balance model of Jones et al. (2007), which was developed for similar steady state mid-latitude lake systems in Turkey as applied in Jones et al. (2007) and Jones and Imbers (2010). We then incorporate the basin geometry (i.e., lake surface area and tributary area), derived from the ArcGIS analysis, to calculate changes in precipitation, relative to modern, at the LGM and post-LGM highstand.

The time-varying (t) water balance and isotopic mass balance for a lake, assuming groundwater flux across the sediment-water interface is negligible, can be described by two equations. The first describes the change in lake volume (V_L) (modified from Jones et al., 2007; Jones and Imbers, 2010; Steinman et al., 2013):

$$\frac{dV_L}{dt} = Q_p + Q_r - Q_e \quad (5)$$

where Q is the input and output fluxes and subscripts correspond to on-lake precipitation (p), runoff (r), and lake surface evaporation (e). The isotopic mass balance is similarly described as (modified from Benson and Paillet, 2002; Jones et al., 2007; Jones and Imbers, 2010; Doebbert et al., 2010; Steinman et al., 2013):

$$\frac{d(\delta^{18}O_L \times V_L)}{dt} = (\delta^{18}O_p \times Q_p) + (\delta^{18}O_r \times Q_r) - (\delta^{18}O_e \times Q_e). \quad (6)$$

where subscripts also apply to the $\delta^{18}O$ of the input and output fluxes. We simplified equation (5) by combining precipitation and runoff into the total surface and subsurface inputs (Q_w) multiplied by the isotopic composition, $\delta^{18}O_w$. Applying the chain rule to equation (6) yields:

$$V_L \frac{d\delta^{18}O_L}{dt} + \delta^{18}O_L \frac{dV_L}{dt} = (\delta^{18}O_w \times Q_w) - (\delta^{18}O_e \times Q_e) \quad (7)$$

Substituting equation (5) into equation (7) and rearranging produces an expression for time varying changes in $\delta^{18}O_L$:

$$V_L \frac{d\delta^{18}O_L}{dt} = ((\delta^{18}O_w - \delta^{18}O_L) \times Q_w) - ((\delta^{18}O_e - \delta^{18}O_L) \times Q_e) \quad (8)$$

Since $Q_e = A_L E_L$ and assuming that for a given lake level, recorded in the isotopic composition and altitude of a tufa sample, $d\delta^{18}O_L/dt = 0$, solving for Q_w yields an expression for the volumetric flux of moisture into the lake:

$$Q_w = \frac{A_L E_L \times (\delta^{18}O_e - \delta^{18}O_L)}{(\delta^{18}O_w - \delta^{18}O_L)} \quad (9)$$

Using the methods and assumptions outlined in Jones et al. (2007), we calculated $\delta^{18}O_e$ and E_L using the equations of Benson and White (1994) for $\delta^{18}O_e$ and the evaporation model of Linacre (1992) for E_L (see **Table 7** for details). The value for $\delta^{18}O_L$ is derived from the measured tufa samples (**Table 3**) and converted using the temperature dependent water-calcite fractionation factor (Kim and O'Neil, 1997) (**Table 7**). Given the

calculation for the total water input flux, Q_w , according to mass balance P_L is calculated assuming that R_T is $0.15 \times P_L$. Integrated watershed modeling of the Lake Estancia basin during the LGM, assuming a colder and wetter climate, suggests that annual runoff (R_T) was 15% of precipitation (P_L) (Menking et al., 2004).

Despite increasing the potential unknowns, the isotope model allows for an explicit solution to a steady-state isotope mass balance for the lake input fluxes (equation (9)), while also considering basin geometry (A_L and A_T). Additionally, we can solve for the basin average precipitation using the isotope model.

5.1.3. Calculating late Pleistocene precipitation amounts for Lake Surprise

To calculate precipitation changes during the LGM and the post-LGM highstand, we apply the hydrologic index (equation (4)) and our isotope-mass balance model (equations (5) through (9)). Our isotope mass balance model also incorporates lake surface area within a basin of fixed total area, thereby decreasing (increasing) tributary area with increasing (decreasing) lake surface area and sample altitude (**Figure A2**). The key parameters used to calculate changes in precipitation are: temperature (T), lake hypsometry (A_L), watershed geometry (A_T), $\delta^{18}O_L$ (from $\delta^{18}O_{calcite}$), $\delta^{18}O_e$, $\delta^{18}O_w$, relative humidity (H_r), and lake surface evaporation rate (E_L) (see **Tables 7** for all values and related equations). We assume that local temperatures were 7 °C lower during the LGM and 5°C lower during the deglacial period, based on pollen assemblages at nearby Little Lake, OR (**Figure 1**; Worona and Whitlock, 2004). Basin hypsometry and watershed geometry was calculated using the ArcGIS analyses described previously. The $\delta^{18}O_L$ is calculated for each sample using the calcite-water fractionation, α_{w-c} , of Kim and O’Neil (1997). The $\delta^{18}O_w$ is assumed to be the average of modern creek waters ($\delta^{18}O_L = -14.57$ ‰ VSMOW; **Table A5**). Calculation of $\delta^{18}O_e$ assumes a kinetic fractionation factor, α_{kin} , for wind speeds <6.8 m/sec of $\alpha_{kin} = 0.994$ (modern average wind speeds are 1.9 m/sec), that the fraction of atmospheric vapor in the lake boundary layer is zero (Benson and White, 1994; Jones et al., 2007), that H_r has not changed and a temperature dependent equilibrium fractionation factor, α_{kin} (Majoube, 1971). Lake surface evaporation rate is calculated using temperature, latitude, lake surface altitude (sample altitude), and modern average wind speed. Jones et al. (2007) note that for similar mid-latitude lakes during the LGM the combined temperature and ice volume effects on incoming precipitation $\delta^{18}O$

would only be between -0.5 ‰ and +1 ‰. Thus, the $\delta^{18}O_w$ (the $\delta^{18}O$ of runoff and precipitation) is held as the average of measured literature values from Surprise Valley (**Tables 7 and A2**; Ingraham and Taylor, 1989; Sladek et al., 2004).

Solving equations (4) and (9) allows us to calculate the absolute and percent change in precipitation for each tufa sample $\delta^{18}O$ measurement, given the associated lake surface altitude and lake surface area for each sample. We do this for LGM samples (n=17) and the post-LGM samples (**Table 8 and Figure 7A**). At the lake highstand, the *HI* calculation predicts an 85.1% increase in precipitation relative to modern, greater than the isotope-mass balance model, which predicts a 75% increase (**Table 8**). This *HI* derived value for the Lake Surprise highstand is similar to those calculated by Reheis (1999) for other proximal lake systems, and Mifflin and Wheat's (1979) prediction of a 77.1% increase for south-central Oregon and northwestern Nevada (**Table 2**). At the LGM, the *HI* calculation predicts a 53.3% increase in precipitation compared to 9.5% increase predicted by the isotope mass balance calculations (**Table 8, Figure 7A and 8**). The primary difference between the two models is that the isotope mass balance calculations allow for sensitivity to changes in humidity and temperature (see equations in **Table 7**). As a result, the ratio of lake precipitation and runoff to net lake surface evaporation, i.e., Q_w/Q_e , is greater than one. Despite the potential uncertainties in the model parameters in **Table 7** and the sensitivity analysis in **Figure A10**, we regard the isotope mass balance calculations as most representative of the calculated precipitation changes at the LGM and post-LGM highstand (**Table 8 and Figure 7A**).

Our calculated changes in precipitation for Surprise Valley are on the lower end of the range of estimates for nearby Lake Lahontan (**Figure 8**) (e.g., Mifflin and Wheat, 1979; Hostetler and Benson, 1990) and for the Basin and Range (e.g., Matsubara and Howard, 2009, Table 1; Menking et al., 2004). Our calculation suggests that decreases in (primarily summer) evaporation (36.4% annual decrease at the LGM, **Table 8 and Figure 8B**), may have played a major role in driving and sustaining lake levels during the late-LGM. Additionally, given the increased lake size and slightly ^{18}O -depleted lake waters during the post-LGM highstand, we find that for a 5°C temperature decrease, precipitation may have increased by up to 75.2% (**Figure 7A**), recording the passing of the polar jet stream (PJS). Other investigations have proposed that the primary driver of

‘pluvial’ lake levels was increased precipitation (e.g., Thompson et al., 1999; Menking et al., 2004; Lyle et al., 2012). Menking et al. (2004) modeled Lake Estancia (34.6°N) and suggest that precipitation doubled at the LGM. Considering the geographic context of Lake Surprise and Lake Estancia, this brackets the ‘mean position’ of the PJS during the LGM (southern Basin and Range, ~35°N), and during the late-HS1 (northern Basin and Range, ~41°N) as proposed by Munroe and Laabs (2012).

5.1.4. Comparison of precipitation estimates to climate model predictions

Predictions of precipitation change between LGM and modern by AOGCMs vary widely among the PMIP3 ensemble members in both magnitude and direction of change (**Table A6**). Evaluation of PMIP3 climate models against paleoclimate data provides an important opportunity to test climate models at mean states very different than modern conditions, such as at the LGM. Thus, direct comparison between climate models and climate records aids in understanding both the past and future regional sensitivity to changing climatic conditions.

Eight of the nine climate models reproduce the modern seasonal precipitation patterns observed at the three meteorological stations in Surprise Valley, while most over-predict absolute precipitation amount (**Figures A11 and A13**). All nine models suggest that the winter contribution to total annual rainfall decreased during the LGM (**Figure 9 and Figures A11 and A13**). The results of the AOGCM analysis (**Table A6**) demonstrate a PMIP3 ensemble average increase of 6.5% LGM precipitation relative to modern, ranging from -14.5 to 50.3% (**Figure 8A and Table A6**). This is comparable to the isotope mass balance calculations, which estimate a 9.0% increase. The predicted total evaporation, a variable parameterized from the surface latent-heat flux by climate models, is predicted to decrease by an ensemble average of 28.2% at the LGM (**Figure 8B and Figure A14**). While not equivalent to lake surface evaporation rate (E_L), these values agree with the 36.4% decrease in evaporation rate estimated for Lake Surprise at the LGM. Among the PMIP3 ensembles, the mean annual temperature LGM anomaly is -7.0°C (**Table A6**), which is the same value used in the isotope mass balance model and predicted by the nearby pollen record (Worona and Whitlock, 2004). Corroborating evidence for temperature control on basin hydrology is the seasonality of peak subsurface and surface runoff predicted by the PMIP3 ensemble. Runoff, which is coupled to

precipitation in the modern, is shifted from late winter in the modern to the late spring during the LGM (**Figure 9** and **Figure A12**). Finally, the PMIP3 ensemble supports our assumption used in the isotope mass balance calculations (**Table 7**) that relative humidity has not changed dramatically since the LGM (only a 4.1% increase).

The climate models and isotope mass balance models presented here provide a suitable framework for investigating whether rainier or cooler conditions (or a balance of the two) drove pluvial lake evolution during the LGM and deglaciation. Based on our analysis, moderate LGM lake levels at Lake Surprise were driven primarily by reduced temperatures and resultant reduced (summer) lake evaporation, with a minimal precipitation increase of 9.5% relative to modern. Following moderate LGM lake levels, Lake Surprise lake levels responded rapidly to a 75.2% increase in precipitation during late HS1. Given the range of uncertainty in some of the input parameters used in the isotope mass balance calculations (see **Figure A10** for the sensitivity analysis of precipitation increase to RH , T , $\delta^{18}O_W$ and wind speed), and the coarse scale of the climate models ($\sim 0.75^\circ$ to 2° horizontal resolution **Figures A13** and **A14**), further constraints are necessary to assess model bias (e.g., climate model downscaling or regional climate modeling, additional isotopic constraints, etc.).

Figure 8. LGM isotope mass balance calculated precipitation (**A**) and lake evaporation (**B**), and the PMIP3 climate model ensemble predictions for Lake Surprise. The values derived by this study are compared to literature values derived using a mass balance model (Miffin and Wheat, 1979) and a thermal evaporation model (Hostetler and Benson, 1990) for Lake Lahontan, and a hydrologic model (Matsubara and Howard, 2009) for the Great Basin. Individual climate model predictions are presented as black dots and a box and whisker plot (**Table A6**). PMIP3 climate models output is total evaporation, calculated from the surface latent heat flux. Isotope mass balance calculations from all LGM samples are presented as a box and whisker plot (**Table 8**). Lake evaporation is calculated using the equations of Linacre (1992) and Jones et al. (2007). See **Figure A10** for the isotope calculation's sensitivity to relative humidity, assumed change in temperature, average wind speed and input $\delta^{18}\text{O}$ (runoff and precipitation). The black and white bars indicate the range of predicted values by the literature.

Figure 9. Seasonal precipitation and runoff predicted by the PMIP3 climate model ensemble. All climate models except COSMOS-ASO uniformly output both precipitation and runoff in the PMIP3 database (**Table A6**). **(A)** PMIP3 ensemble average precipitation normalized to the percent of total annual precipitation for the 0 ka (pre-industrial) and 21 ka (LGM) experiments. Surprise Valley meteorological data is the average of three weather stations (**Figure 2A**). Climate models suggest that the seasonality of precipitation during the LGM was unchanged. **(B)** PMIP3 ensemble average runoff (surface and subsurface) normalized to the percent of total annual runoff for the 0 ka (pre-industrial) and 21 ka (LGM) experiments. Climate models suggest that temperature effects caused the decoupling of precipitation and runoff during the LGM with peak runoff shifting to the late spring. See **Figures A11-14** for individual climate model results.

5.2 Regional implications of the latest Pleistocene Lake Surprise lake-levels

During the LGM, Lake Surprise stood at moderate lake levels (65 to 99 m depth), similar to Lake Lahontan (Benson et al., 1995; Adams et al., 2008), with a *HI* of 0.29 to 0.34 covering 62 to 72% of the ultimate highstand surface area (**Table 2**). During HS1, Lake Surprise rose to its highest and latest Pleistocene lake level (176 m depth) peaking at $>15.23 \pm 0.36$ ka cal BP (^{14}C age), with a concordant ^{230}Th -U isochron age of 18.25 ± 2.67 ka. The *HI* increased by 39% to 0.56, with Lake Surprise covering 36% of the terminally draining watershed. The lack of substantial carbonate deposition and shoreline development (relative to LGM shorelines) suggests this highstand was apparently brief. Recession was subsequently recorded in several stillstands between 15.2 and 12.7 ka. In the following synthesis we link our new MIS 2 Lake Surprise hydrograph (**Figure 7B**), isotope mass balance precipitation calculations (**Figure 7A** and **8**) and PMIP3 ensemble results (**Figure 8** and **9**) to regional records (**Figure 10**) from the LGM through the deglaciation.

Lake and glacial records record the combined effects of changes in temperature and precipitation during the LGM. Coincident with increasing winter insolation and decreasing summer insolation (**Figure 10C**), Lake Franklin and the much larger Lake Bonneville (**Figure 1**) to the east rose gradually during the LGM (**Figure 10G**), although neither reached highstand levels until after the LGM (Oviatt et al., 1992; Munroe and Laabs, 2013). Similarly, statistically discordant radiocarbon ages from similar elevations (1420 to 1454 m) in Lake Surprise record a gradual transgression from 21.82 to 19.13 ka cal BP. At the same time, glaciers were persistent: glacial activity in the southeast Cascades gradually increased through the LGM (**Figure 10F**; Rosenbaum et al., 2012), two glacial advances are recorded in the Sierra Nevada (Tioga 1 and 2) at the beginning and end of the LGM (Phillips et al., 2009), and a glacial maximum is observed at 21.6 ka in the Wallowa Mountains (Licciardi et al., 2004; radiocarbon ages recalculated in Rosenbaum et al., 2012) (**Figure 10E**).

Decreased insolation and reduced temperatures would have decreased summer lake surface evaporation, decreased glacial melt/sublimation, and enhanced basin average runoff. Summer insolation at 40°N was lower than the long-term average between 15.2 and 24.5 ka (during HS1 and the late-LGM), reaching a minimum at 19.7 ka (**Figure**

10I). During the LGM, concurrent increases in regional moisture availability reflected by lake hydrographs and glacial growth, the absence of abrupt shifts in lake and/or glacial records, and decreasing summer insolation, suggest that reduced temperatures decreased lake surface evaporation. The results from our isotope mass balance calculations, and the PMIP3 climate model, support the minimal increase in precipitation (2.5 to 18.2% relative to modern; **Figure 7A** and **8**) and reduced lake surface evaporation (~36% decrease relative to modern; **Figure 8**), corroborating summer insolation as the key driver for moderate LGM lake levels (Munroe and Laabs, 2013; Maher et al., *In Press*). These conditions may have ‘primed’ lake systems to respond rapidly to decreases in North Atlantic temperatures and attendant increases in precipitation (**Figure 10H**) during HS1.

Following the LGM, the temporal correspondence among highstands at Lake Surprise and other small lake systems in the western United States reflects a shift in northern hemisphere mid-latitude atmospheric circulation during HS1. Highstands and significant stillstands of small lake systems during the latter parts of the LGM are attributed to a southward displacement of the ‘mean position’ of the southern arm of the split PJS and ‘dipping westerlies’, which would have brought more precipitation to the region (Negri, 2002; Munroe and Laabs, 2012; Kirby et al., 2013). Our isotope mass balance results require that the pluvial maxima of Lake Surprise was driven by increases in precipitation (~75% increase relative to modern), confirming that displaced mid-latitude storm tracks likely drove increased lake levels during the latter part of HS1. Additionally, reassessment of radiocarbon ages from highstands and significant stillstands spanning 31.8°N to 45.9°N (**Figure 10J**; see **Table A7**) indicates that the regional pluvial maximum lies between ~17 to 15 ka, during the latter part of HS1. Prior to this and immediately following the LGM, during the early part of HS1 (~19 to 17 ka), some authors have suggested a regional ‘Big Dry’ based on desiccation in Lake Estancia, New Mexico (Allen and Anderson, 2000), although this event is not apparent in the lakes in the northern Basin and Range (Broecker et al., 2009; Broecker and Putnam, 2012; Munroe and Laabs, 2013).

Atmospheric forcing of precipitation in the western United States is attributed to North Atlantic cooling during HS1 (COHMAP, 1988; Zic et al., 2002; Denniston et al., 2007; Wagner et al., 2010; Broecker and Putnam, 2012; Munroe and Laabs, 2012;

Benson et al., 1998; 2003; 2013). The proposed mechanism is the winter enhancement of the Aleutian Low via reduced tropical Atlantic precipitation (Okumura et al., 2009) and the suppression of the Atlantic Meridional Overturning Circulation (AMOC) (McManus et al., 2004; Denton et al., 2010). The lake highstand compilation of Munroe and Laabs (2012) demonstrated temporal synchrony between some Basin and Range lake highstands and the onset of iceberg discharge and reduced North Atlantic sea surface temperatures during Heinrich Event 1 at ~17 ka. However, several highstands from smaller lakes near Bonneville and Lahontan post-date this period temporally, as does the Lake Surprise highstand ($>15.23 \pm 0.36$ ka cal BP (^{14}C age)) (**Table A7**). Northern Basin and Range highstands that post-date the apparent mid-HS1 maxima (~17 ka) may have been sustained by regional lake-atmosphere effects (e.g., Hostetler et al., 1994; Licciardi, 2001) brought about by the much larger and longer-lasting Lahontan (highstand at 15.8 ka) (Adams and Wesnousky, 1998) and Bonneville lake systems (recession from Provo shoreline between ~14.5 to 16ka) (Godsey et al., 2011; Miller et al., 2013). Additionally, no significant latitudinal trend (south to north) is captured when considering both highstands and significant stillstands (**Figure 10J**), in contrast to the findings of Lyle et al. (2012). To the south and west respectively, inferences from Lake Elsinore, California (~34°N; Kirby et al., 2013) and Cave of the Bells, Arizona (~32°N, Wagner et al., 2010) provide additional support for the hypothesis that westerly, North Pacific moisture sources were responsible for post-LGM, mid to late HS1, pluvial maximum in the interior western United States. Cumulatively this evidence suggests that the atmospheric augmentation resulted in a latitudinally broad and strengthened mid-latitude westerly storm track during HS1 that was enhanced during Heinrich Event 1 via atmosphere-ocean teleconnections.

Post-LGM glacial activity also supports these conclusions. Although the Laurentide ice-sheet was receding (**Figure 10B**) and northern hemisphere temperature (**Figure 10A** and **10H**) was increasing, persistent glacial activity has been documented. Glacial activity in the southeastern Cascades was recorded until ~15 ka (Rosenbaum et al., 2012; **Figure 10F**), two final post-LGM advances are recognized in the Sierra Nevada (Phillips et al., 2009; **Figure 10E**, Tioga 3 and 4), and a glacial maximum has been observed (~17.4 ka) in the Wallowa Mountains (Licciardi et al., 2004; **Figure 10E**).

Comparable to the small lake systems of similar latitude, these glacial records indicate increased winter precipitation during HS1 despite increasing northern hemisphere temperatures. Finally, while the Laurentide ice-sheet began to retreat at the end of the LGM, the Cordilleran ice sheet advanced southward into parts of Washington, Idaho and Montana until ~ 18 to 15 ka (Waite, 1985; Clague and James, 2002; Dyke, 2004) before fully melting during the Bølling-Allerød and Younger Dryas from 14 to ~11.5 ka (Dyke, 2004). The persistence of the Cordilleran ice sheet may have maintained the southward deflection of the PJS until the onset of the Bølling-Allerød interstadial (~14.5 ka) (COHMAP, 1987; Benson et al., 1990; Garcia and Stokes, 2006; Godsey et al., 2011).

Further evidence for seasonal insolation anomalies driving the moisture balance in the interior western United States is also recorded during the last glaciation via temporal correspondence of a minimum in winter insolation at ~25 ka (**Figure 10I**), speleothem and vein calcite $\delta^{18}\text{O}$ minima (Winograd et al., 2006; Asmerom et al., 2010; Wagner et al., 2010; **Figure 10L** and **10M**), and maximum infiltration rates recorded in soil opal (Maher et al., *In Press*; **Figure 10K**). The minimum in winter insolation may have deepened the Aleutian Low (Ueshima et al., 2006), driving proposed increases in winter precipitation before the LGM (Maher et al., *In Press*). Lakes, soils and speleothems are sensitive to different seasonal variations in the hydrologic cycle. Late-LGM lake levels are particularly sensitive to summer insolation, which in the mid-latitudes is more than twice winter insolation (**Figure 10I**). Given these observations, we propose that changes in orbital conditions that influence seasonal insolation, both summer insolation and its influence on lake evaporation (as well as evapotranspiration), and winter insolation and its influence on precipitation (**Figure 10I**) (Maher et al., *In Press*), are key long-term drivers of hydrologic variability in the western United States.

Figure 10 (next page). Paleoclimate archives and orbital parameters for comparison to the Lake Surprise hydrograph. **(A)** Proxy derived northern hemisphere temperature stack (black line, dashed line is 1σ error bound) deviation from the early Holocene (11.5-6.5 ka) (compiled by Shakun et al., 2012). **(B)** North American ice-sheet percent area deglaciated, including both the Laurentide and Cordilleran ice-sheets (compiled by Dyke, 2004), note the invert scale (gray line). **(C)** Summer (red) and winter (blue) insolation for 40°N (Laskar et al., 2004). **(D)** The Lake Surprise hydrograph (this study) – see the **Figure 7** caption for explanation and symbology. **(E)** Glacier advances (Tioga 1-4) in the Sierra Nevada, CA recorded by cosmogenic ^{36}Cl ages (blue bars; Phillips et al., 2009), and maximal glaciations in the Wallowa Mountains, OR documented by cosmogenic ^{10}Be ages (gray bars; Licciardi et al., 2004) recalculated in Rosenbaum et al. (2012). **(F)** Lacustrine record of glacial flour flux recording glacial extent in the Southeastern Cascades, calculated from the glacial flour content in the Caledonia Marsh core from Upper Klamath Lake (Rosenbaum and Reynolds, 2004; Rosenbaum et al., 2012). **(G)** Lake level curves plotted as the percent of LGM maximum: Lake Bonneville, UT (black dashed line; Oviatt et al., 1992 and others compiled in McGee et al., 2012), Lake Lahontan, NV (black line; Benson et al., 1995; Adams et al., 2008), and Lake Franklin, NV (green line; Munroe and Laabs, 2013). **(H)** NGRIP Greenland ice core $\delta^{18}\text{O}$ on GICC05 timescale (gray line) smoothed with LOESS smoothing function (red line; Rasmussen et al., 2006). **(I)** solar insolation as in **(C)**. **(J)** Lake highstands (black) and stillstands (white boxes) arranged from south to north, as compiled by Munroe and Laabs (2012), Lyle et al. (2012) and recalculated by this study in **Table A7**. **(K)** Great basin and Mojave soil opal P-ET maxima (Maher et al., *In Press*) derived from $(^{234}\text{U}/^{238}\text{U})_0$ variations in soil opal (Maher et al., *In Press*) and vadose zone opal (Paces et al., 2010). **(L)** $\delta^{18}\text{O}$ from Devils Hole, NV (Winograd and others, 2006), interpreted as decreased Pacific sea surface temperatures during $\delta^{18}\text{O}$ minima. **(M)** $\delta^{18}\text{O}$ from speleothem records: Cave of the Bells, AZ (Wager et al., 2010) and Fort Stanton, NM (Asmerom et al., 2010). Both records are interpreted to reflect increased winter precipitation during $\delta^{18}\text{O}$ minima. Both records are smoothed using a LOESS smoothing function (red lines).

6. CONCLUSION

Construction of the late Pleistocene lake hydrograph for Lake Surprise, isotope mass balance calculations and the regional synthesis herein reveals that previously overlooked constraints – namely, the effect of seasonal insolation variability on summer lake evaporation – may have driven lake levels in the Basin and Range during MIS 2. Based on isotope mass balance calculations, moderate LGM lake levels observed in Surprise Valley were driven by decreased evaporation (36.4% relative to modern), with minimal precipitation increases (9.5% relative to modern). These results agree with climate model simulations, which predict a 28% decrease in total evaporation and an average of 6.5% increase in precipitation relative to modern. Evidence presented here questions the long-standing paradigm of a “rainy LGM” in western North America and highlights the strong influence of temperature and seasonal insolation on water availability in the region.

The moderate late-LGM lake levels and sustained glacial activity post-LGM, driven by a minimum in summer insolation (19.7 ka) ‘primed’ the Basin and Range lake systems to respond rapidly to increased rainfall during HS1 brought by latitudinally broad and strengthened mid-latitude westerly storm tracks. This rapid response was recorded in highstands across the Basin and Range (32 to 42°N) via atmosphere-ocean teleconnection to reduced North Atlantic sea surface temperatures (~16.5-17.5 ka until ~14.7 ka). The rapid response of lake systems to substantial increases in precipitation emphasizes the role of atmosphere-ocean teleconnections to western United States water availability. Additionally, recent advances in geochronologic methods (e.g. Maher et al., 2007; Vazquez and Lidzbarski, 2012; Zimmerman et al., 2012; Cheng et al., 2013) will improve the precision necessary to establish global teleconnections. Coordinated studies of runoff/infiltration in small basins (e.g., Maher et al., *In Press*), lake levels (e.g., Garcia and Stokes, 2006; Munroe and Laabs, 2013; *this study*), temperature changes from pollen records (e.g., Worona and Whitlock, 2004) or clumped isotopes (e.g., Lechler et al., 2013), and montane glacial activity (e.g., Rosenbaum et al., 2012), will aid the systematic understanding of spatial and temporal patterns in hydrologic variability of the western United States during MIS 2. Additionally, regional climate modeling (e.g., Ashfaq et al., 2010) with integrated hydrologic models (e.g., Matsubara and Howard, 2009) will better

predict the mean climate and hydrologic states at key time intervals such as the LGM, HS1, Bølling-Allerød and Younger Dryas, and in future warming scenarios (e.g. Quade and Broecker, 2009; Broecker, 2010).

7. PAPER ACKNOWLEDGEMENTS

We thank J. Vazquez (second reader for D. Ibarra's M.S. thesis) and J. Caves for extensive comments and discussions on this manuscript; J. Oster and P. Nuriel for detailed discussions; J. Glen and S. Kraushaar for fieldwork support in Surprise Valley; Guangchao Li for cation analyses; D. Mucciarone for stable isotope analyses; D. Medeiros and C. Kouba from the Stanford Geospatial Center for assistance on the ArcGIS analysis; K. Lau and C. De Massi for help with laboratory work; S. LeRoy for help interpreting the radiocarbon geochronology result; H. Mix for suggesting and discussing the Sr/Ca and stable isotope analyses; and M. Winnick and D. Horton for assistance with the PMIP3 climate model processing. J. Rosenbaum provided the raw data plotted for the glacial flux recorded from the Caledonia Marsh core from Upper Klamath Lake, OR in Figure 10. Lidar data was collected by National Center for Airborne Laser Mapping (NCALM) with funding from NASA through award 10-UAS10-0021 to J. Glen, A. Egger, and C. Ippolito. This work was supported by the National Science Foundation (NSF) grant EAR-0921134 to K. Maher and support from Stanford University.

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9. TABLES

TABLE 1. LOCATION OF SAMPLES

Sample name	Material Type	Sample Group No.	Latitude (°S)	Longitude (°W)	Altitude* (m)	Type of Analysis
<i>Accommodation Zone Shoreline (Figure 2B)</i>						
SVD111-T2	Thick laminated tufa with porous outer rind	1	41.594	120.070	1453.5	U-Th (5 point isochrons (inner rind) + 5 subsamples (outer rind)), ¹⁴ C, Sr/Ca (2 subsamples) and δ18O δ13C (2 subsamples)
SVD111-T3	Thick laminated tufa	1	41.593	120.071	1437.7	U-Th (7 point isochron), ¹⁴ C, Sr/Ca (2 subsamples) and δ18O δ13C (2 subsamples)
SVD111-T4	Thin laminated tufa	1	41.593	120.071	1430.6	U-Th (7 point isochron), ¹⁴ C, Sr/Ca (2 subsamples) and δ18O δ13C (2 subsamples)
SVD111-T14	Thick laminated tufa	2	41.591	120.052	1478.4	U-Th (7 point isochron), ¹⁴ C, Sr/Ca (2 subsamples) and δ18O δ13C (2 subsamples)
SVD111-T18	Densely laminated tufa	3	41.596	120.049	1555.7	U-Th (7 subsamples) ¹⁴ C, Sr/Ca (2 subsamples) and δ18O δ13C
<i>Middle Lake Shoreline Set (Figure 2D)</i>						
SVD112-T1	Thick laminated tufa on porous basalt	1	41.430	119.976	1419.5	U-Th (7 point isochron) and ¹⁴ C, Sr/Ca and δ18O δ13C
SVD112-T2	Laminated tufa	1	41.430	119.976	1419.5	U-Th (7 point isochron and 1.5N HNO ₃ Leach Residue), ¹⁴ C, Sr/Ca (3 replicates) and δ18O δ13C (3 replicates)
SVD112-T3	Thin laminated tufa	1	41.430	119.975	1427.8	U-Th (2 subsamples), Sr/Ca (2 replicates) and δ ¹⁸ O δ ¹³ C (2 replicates)
SVD112-T4	Thin laminated tufa	1	41.430	119.975	1439.0	U-Th (2 subsamples), Sr/Ca (2 replicates) and δ ¹⁸ O δ ¹³ C (2 replicates)
SVD112-T5	Poorly consolidated tufa on porous basalt	1	41.430	119.974	1444.3	U-Th (7 point isochron and 1.5N HNO ₃ Leach Residue), ¹⁴ C, Sr/Ca (2 replicates) and δ18O δ13C (2 replicates)
SVD112-T7	Thin laminated tufa	1	41.428	119.973	1472.5	U-Th (1 sample), Sr/Ca and δ ¹⁸ O δ ¹³ C
SVD112-T9	Thin laminated tufa	2	41.427	119.971	1508.9	U-Th (7 point isochron), ¹⁴ C, Sr/Ca and δ18O δ13C
SVD112-T10	Densely laminated tufa	2	41.427	119.971	1516.8	U-Th (7 point isochron and 1.5N HNO ₃ Leach Residue), ¹⁴ C, Sr/Ca (2 replicates) and δ18O δ13C (2 replicates)
SVD112-T11	Poorly consolidated tufa	3	41.426	119.969	1554.9	U-Th (1 sample), Sr/Ca and δ ¹⁸ O δ ¹³ C
SVD112-T12	Thin laminated tufa	3	41.428	119.968	1576.9	U-Th (7 subsamples), ¹⁴ C, Sr/Ca and δ18O δ13C
<i>Lower Lake Shoreline Set (Figure 2E)</i>						
SVD112-T13	Thick laminated tufa	1	41.217	119.970	1437.2	U-Th (7 point isochron), ¹⁴ C, Sr/Ca and δ18O δ13C
SVD112-T14	Thin laminated tufa	2	41.219	119.965	1530.7	U-Th (7 point isochron), ¹⁴ C, Sr/Ca and δ18O δ13C
<i>Upper Lake Shoreline Set (Figure 2C)</i>						
SVD112-T15	Porous densely consolidated tufa	1	41.718	120.070	1433.1	U-Th (7 point isochron and 1.5N HNO ₃ Leach Residue), ¹⁴ C, Sr/Ca (2 replicates) and δ18O δ13C (2 replicates)
SVD112-T17	Thin laminated tufa	3	41.717	120.062	1542.3	U-Th (1 sample), Sr/Ca and δ ¹⁸ O δ ¹³ C
SVD112-T18	Poorly consolidated tufa	3	41.714	120.062	1564.2	U-Th (2 subsamples), ¹⁴ C, Sr/Ca and δ18O δ13C
SVD112-T19	Poorly consolidated tufa on porous basalt	3	41.712	120.063	1566.8	U-Th (1 sample), Sr/Ca and δ ¹⁸ O δ ¹³ C
<i>Modern Playa Samples (Figure 2A)</i>						
SVD112-C1	Playa Carbonate (Middle Playa Edge)	-	41.533	120.129	-	(²³⁴ U/ ²³⁸ U), (²³⁰ Th/ ²³² Th), Sr/Ca and δ18O δ13C (2 replicates)
SVD112-P1	Playa Surface Sediment (Middle Playa Edge)	-	41.426	119.987	-	1:1 DI Water Extraction and NaOAc Leach: (²³⁴ U/ ²³⁸ U)
SVD112-P2	Playa Surface Sediment (Middle Playa Center)	-	41.533	120.129	-	1:1 DI Water Extraction and NaOAc Leach: (²³⁴ U/ ²³⁸ U)
SVD112-P3	Playa Surface Sediment (Upper Playa)	-	41.602	120.134	-	1:1 DI Water Extraction and NaOAc Leach: (²³⁴ U/ ²³⁸ U)
SVD112-P4	Playa Surface Sediment (Duck Flats)	-	41.079	119.891	-	1:1 DI Water Extraction and NaOAc Leach: (²³⁴ U/ ²³⁸ U)
<i>Modern Water Samples (Figure 2A)</i>						
SVD112-WS1	Stream (Eagle Creek)	-	41.427	119.971	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS2	Stream (Emerson Creek)	-	41.427	119.971	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS3	Stream (Granger Creek)	-	41.473	120.188	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS4	Stream (Deep Creek)	-	41.511	120.215	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS5	Hot Spring (Boyd Spring)	-	41.726	120.083	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS6	Hot Spring (Seyferth Hot Springs)	-	41.614	120.107	-	U concentration and Sr/Ca
SVD112-WS7	Hot Spring (Leonard Hot Springs)	-	41.599	120.092	-	U concentration and Sr/Ca
SVD112-WS8	Hot Spring (Surprise Valley Hot Springs)	-	41.534	120.079	-	U concentration and Sr/Ca
SVD112-WS9	Groundwater (Lake City Well)	-	41.634	120.215	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS10	Stream (Mill Creek)	-	41.643	120.216	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS12	Stream (Fort Bidwell Creek)	-	41.849	120.153	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS13	Pond/Lake (Annie Lake)	-	41.909	120.109	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS15	Groundwater (Cockeral Ranch Well #1)	-	41.658	120.219	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS16	Groundwater (Cockeral Ranch Well #2)	-	41.650	120.215	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS17	Hot Spring (Lake City Hot Springs)	-	41.667	120.212	-	U concentration and Sr/Ca
SVD112-WS18	Hot Spring (Unnamed Hot Spring)	-	41.219	120.065	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS19	Stream (Lost Creek)	-	41.085	119.892	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS20	Groundwater (Eagleville Well)	-	41.323	120.118	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca
SVD112-WS21	Groundwater (Cedarville Well)	-	41.531	120.184	-	U concentration, (²³⁴ U/ ²³⁸ U) and Sr/Ca

*Tufa sample altitude determined by pinning latitude/longitude location to a 0.5m horizontal resolution raster lidar dataset and reported as meters above sea level. Typical elevation error is ± 0.1m (2σ).

TABLE 2. LAKE GEOMETRY, WATERSHED AREA AND HYDROLOGIC INDICES FOR LAKE SURPRISE

Study	Interpreted Age	Lake Elevation (m a.s.l.)	Lake Area, A _L (km ²)	Watershed Area, A _B (km ²)	Watershed Area/Lake Area	Lake Volume (km ³)	Pour Point Elevation (m a.s.l.)	Hydrologic Index (HI)*	Tufa $\delta^{18}\text{O}$ (‰ VPDB)	Δ Temperature (°C)	Estimated Precipitation (mm/yr)	Δ Precipitation [†] (%)
This Study	Modern Meteorology							HI _{met} = 0.11 [†]				
This Study	Modern Playa Lakes	1355 to 1372	0 to 423	3812	> 9.01	0 to 2.1	1621	0 to 0.12				
This Study	MIS 2 (post-LGM)	1531	1366	3812	2.79	164	1621	0.56	-3.845	- 5 [§]	991.0**	75.1**
This Study	MIS 2 (LGM)	1420 to 1440	851 to 977	3812	4.48 to 3.90	39 to 57	1621	0.29 to 0.34	-3.388	- 7 [§]	619.7**	9.5**
This Study	MIS 4 or MIS 6	1567	1499	3812	2.54	218	1621	0.65				
Irwin and Zimbelman (2012)		1545	1430	3837	2.68		1621	0.59				
Zimbelman et al. (2009)		1534										
Orme (2008)			1310	3800	2.90			0.53				
Reheis (1999)	Late Pleistocene	1567	1471	4040	2.75			0.57				
Mifflin and Wheat (1979)	Lahontan	1567	1471	4040	2.75			0.68		- 2.34 ^{††}	552.5 ^{††}	77.1 ^{††}

* HI, (topography derived) hydrologic index = lake surface area/tributary area (basin area minus lake area) (Mifflin and Wheat, 1979; Reheis, 1999)

[†] HI_{met}, (meteorology derived) hydrologic index = runoff from tributary basin/(lake evaporation-lake precipitation) (Mifflin and Wheat, 1979; Reheis, 1999); values from annual average of three modern meteorological stations in Surprise Valley with lake evaporation assumed to be equivalent to pan evaporation. Average precipitation = 566 mm/yr, average pan evaporation = 905 mm/yr and average runoff = 38 mm/yr. Runoff estimated from graphs in Mifflin and Wheat (1979).

[§] Estimated from pollen assemblages at Little Lake, OR (Worona and Whitlock, 1995)

[†] Percent change in precipitation, LGM minus modern. Modern precipitation is taken as 566 mm/yr, as the average of rainfall and snow water equivalent from Lake City (1929-1960), Cedarville (1894-2012) and Fort Bidwell (1911-2011) meteorological stations archived in the Western Regional Climate Center database (<http://www.wrcc.dri.edu/summary/Climsmnca.html>)

** See Table 8 for calculations

^{††} Calculated by Mifflin and Wheat (1979) assuming a 2.34°C (5°F) temperature reduction, 312 mm/yr precipitation and an average hydrologic index of 0.68 in the South Central Oregon and Extreme Northwestern Nevada region

TABLE 3. STABLE ISOTOPE AND ELEMENTAL ANALYSES

Sample name	Sample Group No.	Altitude (m)	1000 × Sr/Ca (mmol/mol)	$\delta^{18}\text{O}$ (‰ VPDB)	$\delta^{13}\text{C}$ (‰ VPDB)	$\delta^{18}\text{O}$ Replicate (‰ VPDB)	$\delta^{13}\text{C}$ Replicate (‰ VPDB)
Accommodation Zone Shoreline Set							
SVDI11-T2-1	1	1453.5	1.070	-3.123	3.866		
SVDI11-T2-2	1	1453.5	1.157	-3.385	3.472		
SVDI11-T3-1a	1	1437.7	1.123	-4.045	3.282		
SVDI11-T3-2	1	1437.7	1.144	-2.988	3.302		
SVDI11-T4-1a	1	1430.6	1.110	-3.238	3.789		
SVDI11-T4-1b	1	1430.6	1.144	-3.225	3.690		
SVDI11-T14-1a	2	1478.4	1.517	-2.501	4.103		
SVDI11-T14-1c	2	1478.4	1.561	-2.851	3.598		
SVDI11-T18-1c	3	1555.7	1.250	-12.559	-6.142		
SVDI11-T18-1e	3	1555.7	0.690				
Middle Playa Shoreline Set							
SVDI12-T1	1	1419.5	1.064	-3.696	3.599		
SVDI12-T2a	1	1419.5	1.273	-3.491	3.504		
SVDI12-T2b	1	1419.5	1.291	-3.472	3.632		
SVDI12-T2c	1	1419.5	1.214	-3.124	3.637		
SVDI12-T3a	1	1427.8	1.304	-3.460	3.618		
SVDI12-T3b	1	1427.8	1.267	-3.457	3.621		
SVDI12-T4a	1	1439.0	1.026	-3.827	3.532		
SVDI12-T4b	1	1439.0	1.039	-3.511	3.555		
SVDI12-T5a	1	1444.3	0.989	-3.888	3.205		
SVDI12-T5b	1	1444.3	1.093	-4.509	2.679		
SVDI12-T7	1	1472.5	1.140	-3.463	3.654	-3.789	3.629
SVDI12-T9	2	1508.9	1.084	-3.587	3.609	-3.465	3.696
SVDI12-T10a	2	1516.8	1.100	-3.552	3.596		
SVDI12-T10b	2	1516.8	1.093	-3.538	3.514		
SVDI12-T11	3	1554.9	1.236	-9.871	-2.032		
SVDI12-T12	3	1576.9	1.115	-9.332	-1.716	-9.434	-1.730
Lower Playa Shoreline Set							
SVDI12-T13	1	1437.2	1.201	-3.343	3.728	-3.387	3.751
SVDI12-T14	2	1530.7	0.878	-3.846	3.737	-3.844	3.702
Upper Playa Shoreline Set							
SVDI12-T15	1	1433.1	1.126	-3.089	3.764		
SVDI12-T15	1	1433.1	1.070	-3.129	3.726		
SVDI12-T17	3	1542.3	0.439	-13.134	-5.565	-13.240	-5.604
SVDI12-T18	3	1564.2	0.645	-11.548	-6.219	-11.789	-6.266
SVDI12-T19	3	1566.8	0.667	-11.041	-6.321	-11.057	-6.402
Modern Carbonate Sample							
SVDI12-C1-1 (Middle Playa Edge)			3.574	-3.998	-0.095	-4.109	-0.131
Sample Group Covariation and Averages							
	$\delta^{18}\text{O}$ - Sr/Ca Covariance (r^2)*			$\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ Covariance (r^2)*	$\delta^{18}\text{O}$ Average	$\delta^{13}\text{C}$ Average	Sr/Ca Average
Sample Group 1	0.06 No Correlation			0.61 Strong	-3.48 ± 0.72	3.56 ± 0.51	1.14 ± 0.18
Sample Group 2	0.92 Strong			0.40 Moderate	-3.33 ± 0.95	3.69 ± 0.39	1.21 ± 0.54
All Samples <1531 m	0.36 Moderate			0.47 Moderate	-3.46 ± 0.78	3.59 ± 0.49	1.16 ± 0.30
Sample Group 3	0.42 Moderate			0.61 Strong	-11.30 ± 2.87	-4.80 ± 4.15	0.86 ± 0.66

* Pearson product-moment coefficient (r^2), rated using no correlation, weak, moderate and strong based on Davis et al. (2009)

TABLE 4. Sr/Ca, [²³⁸U] AND (²³⁴U/²³⁸U) OF MODERN WATER AND PLAYA SAMPLES

Sample name	Material Type	(²³⁴ U/ ²³⁸ U)	[²³⁸ U] (ng/ml)	[Sr] (ng/ml)	[Ca] (µg/ml)	(1000 × Sr)/Ca (mmol/mol)
Modern Water Samples						
SVDI12-WS1	Stream (Eagle Creek)	2.339 ± 0.007	0.0750 ± 0.0021	26.798	8.704	1.408
SVDI12-WS2	Stream (Emerson Creek)	1.846 ± 0.005	0.1039 ± 0.0073	73.097	13.722	2.437
SVDI12-WS3	Stream (Granger Creek)	2.299 ± 0.006	0.4736 ± 0.0124	66.380	24.351	1.247
SVDI12-WS4	Stream (Deep Creek)	2.360 ± 0.006	0.0812 ± 0.0023	107.438	14.630	3.359
SVDI12-WS5	Hot Spring (Boyd Spring)	2.036 ± 0.006	0.2552 ± 0.0069	15.476	2.103	3.366
SVDI12-WS6	Hot Spring (Seyferth Hot Springs)*	-	0.0045 ± 0.0002	548.531	29.163	8.604
SVDI12-WS7	Hot Spring (Leonard Hot Springs)*	-	0.0031 ± 0.0001	236.911	13.973	7.755
SVDI12-WS8	Hot Spring (Surprise Valley Hot Springs)*	-	0.0951 ± 0.0028	260.084	19.437	6.120
SVDI12-WS9	Groundwater (Lake City Well)	2.137 ± 0.006	0.3692 ± 0.0105	181.739	29.890	2.781
SVDI12-WS10	Stream (Mill Creek)	2.045 ± 0.008	0.0262 ± 0.0009	78.503	12.326	2.913
SVDI12-WS12	Stream (Fort Bidwell Creek)	2.399 ± 0.007	0.1137 ± 0.0035	33.688	7.382	2.087
SVDI12-WS13	Pond/Lake (Annie Lake)	1.478 ± 0.005	0.0360 ± 0.0009	30.370	7.677	1.810
SVDI12-WS15	Groundwater (Cockeral Ranch Well #1)	2.130 ± 0.006	0.4574 ± 0.0129	51.796	12.764	1.856
SVDI12-WS16	Groundwater (Cockeral Ranch Well #2)	1.508 ± 0.004	0.9190 ± 0.0535	133.410	25.135	2.428
SVDI12-WS17	Hot Spring (Lake City Hot Springs)*	-	0.0047 ± 0.0003	400.516	22.214	8.247
SVDI12-WS18	Hot Spring (Unnamed Hot Spring)	1.391 ± 0.004	0.0995 ± 0.0029	3.978	2.093	0.869
SVDI12-WS19	Stream (Lost Creek)	1.840 ± 0.005	2.5305 ± 0.1684	102.811	21.788	2.158
SVDI12-WS20	Groundwater (Eagleville Well)	1.797 ± 0.005	0.7240 ± 0.0657	176.296	36.048	2.237
SVDI12-WS21	Groundwater (Cedarville Well)	3.508 ± 0.040	0.9628 ± 0.0815	152.034	29.660	2.345
Modern Playa Samples						
SVDI12-C1	Modern Playa Carbonate (Middle Playa Edge)	1.441 ± 0.005				
Playa Pore Waters						
SVDI12-P1	1:1 DI Water Extraction (Middle Playa Edge)	1.455 ± 0.005				
SVDI12-P2	1:1 DI Water Extraction (Middle Playa Center)	1.640 ± 0.006				
SVDI12-P3	1:1 DI Water Extraction (Upper Playa)	1.595 ± 0.006				
SVDI12-P4	1:1 DI Water Extraction (Duck Flats)	1.591 ± 0.006				
Playa Authigenic Carbonate Fraction						
SVDI12-P1	NaOAc Carbonate Leach (Middle Playa Edge)	1.327 ± 0.004				
SVDI12-P2	NaOAc Carbonate Leach (Middle Playa Center)	1.542 ± 0.006				
SVDI12-P3	NaOAc Carbonate Leach (Upper Playa)	1.513 ± 0.004				
SVDI12-P4	NaOAc Carbonate Leach (Duck Flats)	1.571 ± 0.004				

* (²³⁴U/²³⁸U) not measured due to low U concentrations

TABLE 5. ISOCHRON DERIVED ACTIVITY RATIOS AND DETRITAL THORIUM CORRECTED AGES																		
TSD Rosholt Isochron Carbonate Activity Ratios (All Sub-Samples)		TSD Method Detrital Corrected Age (All Sub-Samples) 2s Rosholt Isochron Derived				TSD Rosholt Isochron Carbonate Activity Ratios (Most Robust Fit) [†]		TSD Method Detrital Corrected Age (Most Robust Fit) 2s Rosholt Isochron Derived				TSD Detrital Corrected Age (MLE Osmond 3-D Solution)				Single Sample Correction Error-Weighted Avg		Radiocarbon Age
Sample name	No. of Sub-Samples	^(230Th/232U) *	^(230Th/232U) †	Corrected Age (ka BP) [‡]	^(230Th/232U) total	No. of Sub-Samples	^(230Th/232U) *	^(230Th/232U) †	Corrected Age (ka cal BP) [‡]	^(230Th/232U) total	No. of Sub-Samples	^(230Th/232U)	^(230Th/232U)	Corrected Age (ka BP) [‡]	^(230Th/232U) total	Corrected Age (ka cal BP) [‡]		
Accommodation Zone Shoreline Set																		
SVD11-T2 Outer Rind	5	No Robust Rosholt (^(230Th/232Th) vs (^(230Th/232Th)) Isochron				No Robust Rosholt (^(230Th/232Th) vs (^(230Th/232Th)) Isochron				No Robust Osmond Isochron								
SVD11-T2 Inner Rind	5	0.283 ± 0.051	1.561 ± 0.049	21.52 ± 4.34	1.597 ± 0.051	4	0.289 ± 0.014	1.563 ± 0.064	21.93 ± 1.53	1.599 ± 0.066	4	0.285 ± 0.026	1.560 ± 0.029	21.7 ± 2.2	1.595 ± 0.031	17.28 ± 1.61	19.13 ± 0.26	
SVD11-T3	7	0.325 ± 0.058	1.534 ± 0.044	25.52 ± 5.14	1.574 ± 0.047	4	0.337 ± 0.014	1.529 ± 0.054	26.69 ± 1.63	1.571 ± 0.057	4	0.335 ± 0.023	1.530 ± 0.024	26.5 ± 2.1	1.572 ± 0.026	21.47 ± 0.94	20.92 ± 0.41	
SVD11-T4	7	0.360 ± 0.125	1.618 ± 0.076	26.97 ± 10.62	1.667 ± 0.082	4	0.254 ± 0.022	1.584 ± 0.087	18.84 ± 2.08	1.616 ± 0.089	4	0.247 ± 0.031	1.603 ± 0.034	18.1 ± 2.6	1.635 ± 0.034	11.41 ± 1.87	20.42 ± 0.19	
SVD11-T14	7	0.207 ± 0.155	1.983 ± 0.212	11.82 ± 9.41	2.017 ± 0.217	7	0.207 ± 0.155	1.983 ± 0.212	11.82 ± 9.41	2.017 ± 0.217	No Robust Osmond Isochron			11.82 ± 2.07	1.682 ± 0.247	12.70 ± 0.13		
SVD11-T18	7	0.836 ± 1.263	1.304 ± 0.075	105.84 ± 258.29	1.409 ± 0.311	No Robust Rosholt (^(230Th/232Th) vs (^(230Th/232Th)) Isochron				No Robust Osmond Isochron				15.41 ± 5.85	1.261 ± 0.009	16.33 ± 0.47		
Middle Lake Shoreline Set																		
SVD12-T1	7	0.343 ± 0.066	1.567 ± 0.072	26.44 ± 5.86	1.612 ± 0.076	7	0.343 ± 0.066	1.567 ± 0.072	26.44 ± 5.86	1.612 ± 0.076	7	0.342 ± 0.048	1.555 ± 0.075	26.6 ± 4.8	1.599 ± 0.075	18.54 ± 0.88	20.91 ± 0.40	
SVD12-T2	7 + Residue	0.311 ± 0.015	1.624 ± 0.016	22.78 ± 1.23	1.666 ± 0.017	7 + Residue	0.311 ± 0.015	1.624 ± 0.016	22.78 ± 1.23	1.666 ± 0.017	7 + Residue	0.312 ± 0.014	1.624 ± 0.019	23.0 ± 1.3	1.665 ± 0.018	24.84 ± 0.92	21.82 ± 0.34	
SVD12-T3	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	1.544 ± 0.003		
SVD12-T4	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	1.527 ± 0.104		
SVD12-T5	7 + Residue	0.306 ± 0.120	1.611 ± 0.060	22.57 ± 9.81	1.651 ± 0.064	5 + Residue	0.310 ± 0.050	1.615 ± 0.064	22.87 ± 4.17	1.656 ± 0.067	5 + Residue	0.310 ± 0.057	1.600 ± 0.055	23.2 ± 5.2	1.640 ± 0.054	19.80 ± 2.00	10.69 ± 0.11	
SVD12-T7	1	0.355 ± 0.154	2.064 ± 0.209	20.18 ± 9.74	2.126 ± 0.216	5	0.190 ± 0.056	1.803 ± 0.356	12.01 ± 4.50	1.831 ± 0.363	No Robust Osmond Isochron			11.78 ± 2.85	1.840 ± 0.021	14.55 ± 0.42		
SVD12-T9	7	0.271 ± 0.025	1.867 ± 0.126	16.83 ± 2.10	1.909 ± 0.130	5	0.269 ± 0.010	1.888 ± 0.062	16.48 ± 0.87	1.931 ± 0.063	5	0.268 ± 0.021	1.886 ± 0.090	16.5 ± 1.7	1.929 ± 0.092	17.45 ± 0.47	14.85 ± 0.35	
SVD12-T10	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	1.507 ± 0.039		
SVD12-T11	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	1.543 ± 0.011		
SVD12-T12	7	No Robust Rosholt (^(230Th/232Th) vs (^(230Th/232Th)) Isochron				No Robust Rosholt (^(230Th/232Th) vs (^(230Th/232Th)) Isochron				No Robust Osmond Isochron				29.21 ± 8.45	1.197 ± 0.024	16.67 ± 6.57	8.58 ± 0.07	
Lower Lake Shoreline Set																		
SVD12-T13	7	0.274 ± 0.018	1.554 ± 0.073	20.85 ± 1.84	1.587 ± 0.076	7	0.274 ± 0.018	1.554 ± 0.073	20.85 ± 1.84	1.587 ± 0.076	7	0.273 ± 0.039	1.581 ± 0.044	20.9 ± 3.4	1.581 ± 0.044	21.30 ± 1.13	20.85 ± 0.42	
SVD12-T14	7	0.278 ± 0.104	1.789 ± 0.060	18.16 ± 7.35	1.831 ± 0.064	5	0.279 ± 0.037	1.785 ± 0.060	18.25 ± 2.67	1.827 ± 0.062	5	0.274 ± 0.035	1.803 ± 0.045	17.8 ± 2.5	1.845 ± 0.046	18.98 ± 2.14	15.23 ± 0.36	
Upper Lake Shoreline Set																		
SVD12-T15	7	0.290 ± 0.099	1.576 ± 0.052	21.84 ± 8.19	1.612 ± 0.056	5	0.260 ± 0.016	1.575 ± 0.054	19.44 ± 1.46	1.607 ± 0.056	5	0.260 ± 0.044	1.572 ± 0.039	19.5 ± 3.6	1.605 ± 0.040	19.71 ± 1.12	19.32 ± 0.13	
SVD12-T17	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	1.996 ± 0.003		
SVD12-T18	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	1.128 ± 0.004		
SVD12-T19	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	24.70 ± 2.25	25.00 ± 0.45	

* Slope of Rosholt Isochron ($^{230}\text{Th}/^{232}\text{Th}$) vs ($^{230}\text{Th}/^{232}\text{Th}$)

† Slope of Rosholt Isochron ($^{230}\text{Th}/^{232}\text{Th}$) vs ($^{230}\text{Th}/^{232}\text{Th}$)

‡ BP stands for "Before Present" where the "Present" is defined as the year 1950 A.D.

* See Table A3 (footnote *) for samples omitted from robust fit Rosholt Isochrons

** See Table A3 (footnotes † and §) for samples omitted from error-weighted mean

TABLE 6. NEW RADIOCARBON AGES FOR LAKE SURPRISE

Sample name	Laboratory Number	Altitude (m)	¹⁴ C age (yr) ± 1σ	Calibrated age range (yr cal. BP)* [†] ± 2σ	Relative area under distribution	Median Age (yr cal BP) [‡]	Calibrated age (ka cal. BP) ^{§,¶} ± 2σ IntCal09
Accommodation Zone Shoreline Set							
SVDI11-T2-1 Inner Rind	Beta - 342012	1453.5	15930 ± 70	18871 - 19387	(1.000)	19126	19.13 ± 0.26
SVDI11-T3-2	Beta - 342013	1437.7	17580 ± 70	20513 - 21327	(1.000)	20950	20.92 ± 0.41
SVDI11-T4-1b	Beta - 340102	1430.6	17280 ± 60	20230 - 20604	(0.644)	20493	20.42 ± 0.19
				20637 - 21057	(0.356)		20.85 ± 0.21
SVDI11-T14-1c	Beta - 342014	1478.4	10790 ± 50	12572 - 12831	(1.000)	12670	12.70 ± 0.13
SVDI11-T18-1c	Beta - 340103	1555.7	13310 ± 40	15632 - 15745	(0.023)	16413	15.69 ± 0.06
				15862 - 16804	(0.977)		16.33 ± 0.47
Middle Lake Shoreline Set							
SVDI12-T1-a	Beta - 340104	1419.5	17560 ± 60	20507 - 21303	(1.000)	20919	20.91 ± 0.40
SVDI12-T2-b	Beta - 342015	1419.5	18270 ± 70	21480 - 22156	(1.000)	21815	21.82 ± 0.34
SVDI12-T5-b	Beta - 342016	1444.3	9470 ± 40	10581 - 10793	(0.883)	10713	10.69 ± 0.11
				10854 - 10861	(0.004)		10.86 ± 0.00
				10962 - 11006	(0.055)		10.98 ± 0.02
				11023 - 11065	(0.058)		11.04 ± 0.02
SVDI12-T9-1	Beta - 340105	1508.9	12420 ± 50	14129 - 14967	(1.000)	14486	14.55 ± 0.42
SVDI12-T10-b	Beta - 342017	1516.8	12600 ± 50	14266 - 14320	(0.014)	14878	14.29 ± 0.03
				14500 - 15199	(0.986)		14.85 ± 0.35
SVDI12-T12-1	Beta - 340106	1576.9	7810 ± 40	8457 - 8498	(0.045)	8587	8.48 ± 0.02
				8508 - 8656	(0.925)		8.58 ± 0.07
				8670 - 8699	(0.030)		8.68 ± 0.01
Lower Lake Shoreline Set							
SVDI12-T13	Beta - 342018	1437.2	17490 ± 90	20426 - 21274	(1.000)	20820	20.85 ± 0.42
SVDI12-T14	Beta - 342019	1530.7	12750 ± 50	14770 - 14830	(0.015)	15127	14.80 ± 0.03
				14871 - 15583	(0.985)		15.23 ± 0.36
Upper Lake Shoreline Set							
SVDI12-T15-b	Beta - 342020	1433.1	16150 ± 70	18937 - 19179	(0.334)	19298	19.06 ± 0.12
				19184 - 19451	(0.666)		19.32 ± 0.13
SVDI12-T18	Beta - 342021	1564.2	20970 ± 110	24552 - 25453	(1.000)	25002	25.00 ± 0.45

* Calibrated using the Calib 6.0 program with IntCal09 (Stuiver and Reimer, 1993; Reimer et al., 2009). See supplementary information (Figures A6-9) for calibration plots.

[†] BP stands for "Before Present" where the "Present" is defined as the year 1950 A.D.

[‡] Median age calculated using the Calib 6.0 program

[§] Bolded ages are the preferred ages with the highest probability and are used for comparison to the ²³⁰Th-U ages (Table 5, Figure 5)

TABLE 7. ISOTOPE MODEL PARAMETER VALUES

Parameters	Description	Value
T	Temperature (°C)	Modern = 9.2, LGM = 2.2, post-LGM = 4.2 based on Worona and Whitlock (2004)
P _L	Precipitation (L, lake) (mm/yr)	Modern = 566 mm/yr (solved by model)
R _T	Runoff (surface+sub-surface) (mm/yr)	Unknown (assumed $0.15 \times P_T$, based on Menking et al. (2004))
H _r	Relative Humidity	Unknown (Modern = 0.57%)
E _L	Lake surface evaporation rate (mm/yr)	Linacre (1992) based on T, latitude (41.5°N), lake surface altitude and wind speed (modern=1.9 m/s)*
A _L	Lake Area (km ²)	Varies with lake surface level - based on modern topography, calculated using fifth order polynomial fit through results of DEM analysis [†]
A _B	Basin Area (km ²)	3812 km ² based on modern topography (DEM analysis)
A _T	Tributary Area (km ²)	$A_T = A_B - A_L$
$\delta^{18}O_L$	Lake water $\delta^{18}O$	Calculated using α_{w-c} at given T from tufa sample measurements
$\delta^{18}O_{in}$	Input Fluxes (Precipitation+ Runoff) $\delta^{18}O$	-14.57‰ (VSMOW, modern creek average Table DR5)
$\delta^{18}O_e$	Evaporation $\delta^{18}O$	Benson and White (1994) assuming fraction of atmospheric water vapor in lake boundary layer, f_{ad} is negligible ($f_{ad} = 0$) [‡]
Q _w	Total Input Flux	$Q_p + Q_r$
Q _p	Lake Surface Precipitation Flux	$P_L \times A_L$
Q _r	Runoff Flux	$R_T \times A_T$
Q _e	Evaporation Flux	$E_L \times A_L$
α_{w-c}	Calcite-water fractionation	Kim and O'Neil (1997) [§]
α_{kin}	Kinetic evaporation fractionation	0.994 for wind speeds <6.8 m/s, Majoube (1971)
α_{eq}	Equilibrium evaporation fractionation	Majoube (1971)**

* $E_L = [0.015 + 4 \times 10^{-4}T + 10^{-6}z] \times [480(T + 0.006z) / (84 - Lat) - 40 + 2.3u(T - T_d)]$, where T is mean annual temperature, z is lake altitude (meters a.s.l.), u is wind speed, Lat is latitude and T_d is dew point temperature. T_d (°C) is calculated as: $T_d = 0.52 \times T_{min} + 0.6 \times T_{max} - (0.009 \times (T_{max})^2) - 2$. T_{min} and T_{max} are assumed to uniformly shift by the same magnitude as mean annual temperature (T). (Linacre, 1992; Jones et al., 2007)

[†] $A_L = (-25.028) + (44.127 \times h_L) + ((-1.0357) \times h_L^2) + (0.013333 \times h_L^3) + ((-0.000088611) \times h_L^4) + (0.00000029112 \times h_L^5) + ((-0.0000000037455) \times h_L^6)$ where h_L = lake surface altitude (altitude minus 1355 meters).

[‡] $\delta^{18}O_e = (R_{evap} - 1) \times 1000$, where $R_{evap} = R_{lake} \times (\alpha_{kin}/\alpha_{eq}) \times (1/(1 - H_r + (H_r \times \alpha_{kin})))$ and $R_{lake} = (\delta^{18}O_L / 1000) + 1$. (Benson and White, 1994)

[§] $1000 \times \ln(\alpha_{calcite-water}) = 18.03 \times (10^3 \times T^{-1}) - 32.42$. T in Kelvin, assumed to be average summer temperature uniformly shifted by the same magnitude as mean annual temperature (T). (Kim and O'Neil, 1997)

** $\alpha_{eq} = \exp(1137 \times T^{-2} - 0.4156 \times T^{-1} - 0.0020667)$. T in Kelvin. (Majoube, 1971)

TABLE 8. HYDROLOGIC INDEX (HI) AND ISOTOPE MASS BALANCE CALCULATIONS

Hydrologic Index (HI) Calculation*								Isotope Mass Balance Calculation†						
	Lake Level Altitude	Water Depth	Surface Area	Tributary Area	Hydrologic Index	Precipitation	Δ P	δ ¹⁸ O _{calcite}	δ ¹⁸ O _L	Lake Evaporation	δ ¹⁸ O _e	Q _w	Precipitation	Δ P
Sample	(m a.s.l.)	(m)	(km ²)	(km ²)		(mm/yr)	(%)	(‰ VPDB)	(‰ VSMOW)	(mm/yr)	(‰ VSMOW)	(m ³ /yr)	(mm/yr)	(%)
LGM Samples														
SVDI11-T2-1	1453.5	98.5	1031	2781	0.371	917.5	62.1	-3.123	-3.448	585.5	-20.64	932804690	644.2	13.9
SVDI11-T2-2	1453.5	98.5	1031	2781	0.371	917.5	62.1	-3.385	-3.710	585.5	-20.90	955057055	659.6	16.6
SVDI11-T3-1a	1437.7	82.7	944	2868	0.329	879.1	55.4	-4.045	-4.369	578.1	-21.54	918508324	668.6	18.2
SVDI11-T3-2	1437.7	82.7	944	2868	0.329	879.1	55.4	-2.988	-3.313	578.1	-20.51	833179570	606.5	7.2
SVDI11-T4-1a	1430.6	75.6	907	2905	0.312	861.9	52.3	-3.238	-3.563	574.9	-20.75	814290165	606.5	7.2
SVDI11-T4-1b	1430.6	75.6	907	2905	0.312	861.9	52.3	-3.225	-3.550	574.9	-20.74	813284920	605.7	7.0
SVDI12-T1	1419.5	64.5	854	2957	0.289	835.8	47.7	-3.696	-4.020	569.7	-21.20	792886253	610.8	8.0
SVDI12-T2a	1419.5	64.5	854	2957	0.289	835.8	47.7	-3.491	-3.816	569.7	-21.00	777951961	599.3	5.9
SVDI12-T2b	1419.5	64.5	854	2957	0.289	835.8	47.7	-3.472	-3.796	569.7	-20.98	776575497	598.3	5.7
SVDI12-T2c	1419.5	64.5	854	2957	0.289	835.8	47.7	-3.124	-3.449	569.7	-20.64	752560774	579.8	2.5
SVDI12-T3a	1427.8	72.8	893	2918	0.306	855.2	51.1	-3.460	-3.785	573.6	-20.97	816361203	613.3	8.4
SVDI12-T3b	1427.8	72.8	893	2918	0.306	855.2	51.1	-3.457	-3.782	573.6	-20.97	816126318	613.2	8.4
SVDI12-T4a	1439	84	950	2861	0.332	882.2	55.9	-3.827	-4.152	578.7	-21.33	907069092	657.5	16.2
SVDI12-T4b	1439	84	950	2861	0.332	882.2	55.9	-3.511	-3.836	578.7	-21.02	880661324	638.3	12.8
SVDI12-T13	1437.2	82.2	941	2871	0.328	877.8	55.1	-3.343	-3.668	577.9	-20.86	857278731	625.1	10.5
SVDI12-T15	1433.1	78.1	920	2892	0.318	867.9	53.4	-3.089	-3.414	576.0	-20.61	816317153	603.1	6.6
SVDI12-T15	1433.1	78.1	920	2892	0.318	867.9	53.4	-3.129	-3.454	576.0	-20.65	819220785	605.3	7.0
LGM Average Precip (HI):						867.6	53.3	LGM Average Lake Evap (Isotope Model):		575.9	LGM Average Precip (Isotope Model):		619.7	9.5
Post-LGM Samples														
SVDI12-T7	1472.5	117.5	1134	2678	0.424	959.6	69.6	-3.463	-3.350	776	-20.34	1332561746	867.7	53.3
SVDI12-T7	1472.5	117.5	1134	2678	0.424	959.6	69.6	-3.789	-3.675	776	-20.66	1371919912	893.3	57.9
SVDI11-T14-1a	1478.4	123.4	1163	2648	0.439	971.2	71.6	-2.501	-2.387	779	-19.40	1264856983	810.5	43.2
SVDI11-T14-1c	1478.4	123.4	1163	2648	0.439	971.2	71.6	-2.851	-2.737	779	-19.74	1301799929	834.1	47.4
SVDI12-T9	1508.9	153.9	1284	2528	0.508	1019.0	80.1	-3.587	-3.473	794	-20.46	1560352962	938.3	65.8
SVDI12-T9	1508.9	153.9	1284	2528	0.508	1019.0	80.1	-3.465	-3.352	794	-20.34	1543646580	928.2	64.0
SVDI12-T10a	1516.8	161.8	1309	2503	0.523	1029.3	81.9	-3.552	-3.438	798	-20.43	1593901125	946.2	67.2
SVDI12-T10b	1516.8	161.8	1309	2503	0.523	1029.3	81.9	-3.538	-3.424	798	-20.42	1591921357	945.0	67.0
SVDI12-T14	1530.7	175.7	1354	2458	0.551	1047.2	85.1	-3.846	-3.733	805	-20.72	1089307483	991.1	75.2
SVDI12-T14	1530.7	175.7	1354	2458	0.551	1047.2	85.1	-3.844	-3.730	805	-20.72	1089307483	990.9	75.1

* Assumes no change in temperature or humidity relative to modern

† See Table 7 for isotope mass balance calculation assumption

Appendix A

Supplemental figures and tables

Appendix A contains fourteen supplemental figures and seven supplemental tables supporting the methods, results and discussion of this study. Figure captions and legends are given with each figure. Appendix A was submitted as supplemental data repository file for submission of Chapter 1 to *Geological Society of America Bulletin*.

A1. Supplementary Figures

Figure A1. Long-term stability of U and Th standard measurements at the Stanford ICP-MS/TIMS Facility using a Nu Instruments Plasma HR multi-collector inductively coupled plasma mass spectrometer (MC-ICP-MS). The number of measurements is given only when multiple measurements were analyzed on a given day. Stanford measurements (single measurements and daily averages) are shown as open white circles, with the long term Stanford average shown as a black filled circle. All averages are un-weighted averages. Recent literature values are given as black squares. (A) U activity standard SRM 4321A measurements are compared to

values reported by Oster et al. (2012) and Neymark (2011) (who report measurements on a different aliquot, SRM 4321B). **(B)** U isotope standard CRM-145 (also known as NBL-112A) measurements are compared to the CRM 145 certified value, Oster et al. (2012), Shen et al. (2012), Torfstein et al. (2012), Placzek et al. (2006) and Cheng et al. (2000). **(C)** Th isotope standard OU Th ‘U’ measurements are compared to the Sims et al. (2008) working group average, Turner et al. (2001), Turner et al. (1997) and Thomas et al. (1999). **(D)** Th isotope standard IRMM-036 measurements are compared to Neymark (2011), Sims et al. (2008) working group average and the IRMM-036 certified value (procedure for gravimetric preparation in Raptis et al., 1998).

Figure A2. Determination of Lake Surprise hypsometric data using modern topography. **(A)** Using the ArcGIS 10.1 hydrology toolbox, the USGS National Elevation Dataset merged DEM was analyzed to calculate the basin size, lake surface area and volume. Lake surface contours (green) were created at 10 meter intervals. **(B)** Lake volume (km³), lake surface area (km²) and hydrologic index (HI) vs. lake level (meters) and elevation (meters a.s.l.). The LGM lake levels and post-LGM highstand are shown as grey bars.

Figure A3 (Previous Page). (A-D) Isochron plots of coeval samples for the Accommodation Zone tufa samples. All error ellipses and error envelopes are 2σ . (i-ii) “Rosholt” isochrons of $(^{230}\text{Th}/^{232}\text{Th})$ vs. $(^{238}\text{U}/^{232}\text{Th})$ and $(^{234}\text{U}/^{232}\text{Th})$ vs. $(^{238}\text{U}/^{232}\text{Th})$ (measured values), with error-weighted linear regressions (see text for details), calculates the $(^{230}\text{Th}/^{238}\text{U})_{\text{authigenic}}$ and $(^{234}\text{U}/^{238}\text{U})_{\text{authigenic}}$, based on the slopes of the linear regressions. The intercept of $(^{230}\text{Th}/^{232}\text{Th})$ vs. $(^{238}\text{U}/^{232}\text{Th})$ is the $(^{230}\text{Th}/^{232}\text{Th})_{\text{initial}}$ of the suite of coeval samples. Samples not included in linear regressions are grey ellipses. (iii)

Projected representation of a 3-D “Osmond” isochron calculation of ^{230}Th -U ages for coeval suites of samples (see text for details) onto a ^{230}Th - ^{234}U - ^{238}U evolution plot. Black ellipses are measured values and white ellipses are projected values based on the projected isochron (grey line). Grey ellipses are the calculated ^{230}Th -U age and $(^{234}\text{U}/^{238}\text{U})_0$ of the sample.

Figure A4 (Above). Isochron plots of coeval samples for the Lower Lake (A-B) and Upper Lake (C) tufa samples. See **Figure A3** caption and legend for description of symbology and calculations.

Figure A5 (Next Page). (A-E) Isochron plots of coeval samples for the Middle Lake tufa samples. See **Figure A3** caption and legend for description of symbology and calculations.

Figure A6. Calibration of Accommodation Zone carbonate samples' radiocarbon (^{14}C) results to calendar years BP. Calibration was carried out using the extended IntCal09 calibration curve using Calib 6.0 program (Stuiver and Reimer, 1993; Reimer et al., 2009).

Figure A7. Calibration of Middle Lake carbonate samples' radiocarbon (^{14}C) results to calendar years BP. Calibration was carried out using the extended IntCal09 calibration curve using Calib 6.0 program (Stuiver and Reimer, 1993; Reimer et al., 2009).

Figure A8. Calibration of Lower Lake carbonate samples' radiocarbon (^{14}C) results to calendar years BP. Calibration was carried out using the extended IntCal09 calibration curve using Calib 6.0 program (Stuiver and Reimer, 1993; Reimer et al., 2009).

Figure A9. Calibration of Upper Lake carbonate samples' radiocarbon (^{14}C) results to calendar years BP. Calibration was carried out using the extended IntCal09 calibration curve using Calib 6.0 program (Stuiver and Reimer, 1993; Reimer et al., 2009).

Figure A10. Sensitivity analysis of the isotope mass balance calculations for LGM precipitation increases (Tables 10 and 11; Figure 10; Equations 5-9) for an average LGM tufa sample $\delta^{18}\text{O}_{\text{VPDB}}$ (-3.34‰) and lake level (77.9 meters depth). Assuming all values listed in Table 10 relative humidity (A), mean annual and summer average temperature change (modern-LGM) (B), input $\delta^{18}\text{O}_{\text{VSMOW}}$ (precipitation and runoff) (C), and average wind speed (D) are varied separately within ranges of modern variability. The values assumed by our mass balance calculations (Table 10) are shown as black dots. (E) The range of values calculated for all LGM tufa samples (Table 11) is shown as a box and whisker plot (with a median and interquartile range), and is shaded in grey (B-D).

Figure A11. Normalization of raw climate model output precipitation amounts (**A and C**) from bilinear interpolation to 120°W, 41.5°N for preindustrial (**A-B**) and 21 ka (**C-D**) simulations. Due to the large spread in absolute precipitation values, models from the PMIP3 ensemble were normalized to the percent of annual precipitation (**B and D**) for comparison. All models except CNRM-CM5 appear to re-produce seasonality in the pre-industrial climate model runs (compared to the averaged meteorological station data).

Figure A12. Normalization of raw climate model output runoff amounts (**A and C**) from bilinear interpolation to 120°W, 41.5°N for preindustrial (**A-B**) and 21 ka (**C-D**) simulations. Due to the large spread in absolute runoff values, models from the PMIP3 ensemble were normalized to the percent of annual runoff (**B and D**) for comparison. COSMOS-ASO did not archive runoff in the PMIP3 database.

Figure A13. Maps of the western United States showing the annual LGM precipitation anomaly for the nine PMIP3 climate models used in this study. Values are calculated as the absolute difference between the 21 ka (LGM) and 0 ka (pre-industrial) experiment precipitation amounts. Note that the color bars are not the same scale. No bias correction was applied and all maps were made using the original resolution of the climate model outputs. The location of Lake Surprise is given as an open black star.

Figure A14. Maps of the western United States showing the annual LGM infiltration (P-ET) anomaly for the nine PMIP3 climate models used in this study. Values are calculated as the absolute difference between the 21 ka (LGM) and 0 ka (pre-industrial) experiment infiltration amounts. Infiltration (q) is calculated as: $q = P - ET$. Where P is precipitation and ET is the total evaporation. Note that the color bars are not the same scale. No bias correction was applied and all maps were made using the original resolution of the climate model outputs. The location of Lake Surprise is given as an open black star.

A2. Supplementary Tables

TABLE A1. PREVIOUS LITERATURE CONSTRAINTS ON LAKE SURPRISE LAKE LEVEL

Elevation (m a.s.l.)	Age Constraints	Dating Method*	Details	Location	Reference
~1378 m	23.2 ± 0.2 (ka cal BP)	Radiocarbon	Trego Hot Spring (THS) tephra moderate to deep-water deposition Correlated tephra, age from Pyramid Lake (Benson et al., 1997)	Playa edge in two locations, see Figure 2A	Hedel (1980; 1984)
1475 and 1493 m	~ 18 to 24 ka 18.40 ± 3.17 (ka) 28.08 ± 4.44 (ka)	OSL + IRSL [†] IRSL	Shallow water deltaic deposits post-dating THS Youngest sub-lacustrine age on deltaic deposits Oldest sub-lacustrine age on deltaic deposits, age questionable due to plagioclase contamination	Cooks Canyon Trench Cooks Canyon Trench	Personius et al. (2009)
1475 and 1493 m	1 to 13 ka 1.24 ± 0.07 (ka cal BP) 1.15 ± 0.26 (ka cal BP) 7.86 ± 0.12 (ka cal BP) 13.38 ± 2.64 (ka)	Radiocarbon Radiocarbon Radiocarbon OSL + IRSL [†]	Subaerial sediments with Mazama Ash Youngest radiocarbon age on charcoal from buried soil Youngest Radiocarbon age on charred plant tissue Oldest radiocarbon age on charcoal found directly below Mazama Ash Oldest sub-aerial age on post-lacustrine eolian sand	Steamboat Canyon Cooks Canyon Trench Cooks Canyon Trench Cooks Canyon Trench	Personius et al. (2009)
~1355 m	Younger than ~ 6 ka 5.64 ± 0.155 yr BP (¹⁴ C age)	Radiocarbon	Prehistoric settlements at the edge of the playa Age from bone found in the second deepest house	King's Dog archeological site	O'Connell and Inoway (1994)

* OSL - blue light optically stimulated luminescence; IRSL - infrared-stimulated luminescence

[†] Error-weighted mean

TABLE A2. U-SERIES ISOTOPE RATIOS AND CONCENTRATIONS FOR ROCK STANDARDS AND ISOTOPIC STANDARDS MEASURED DURING SAMPLE ANALYSES

	No. U Analyses	No. Th Analyses	Repl. diss.	(²³⁴ U/ ²³⁸ U)	(²³⁰ Th/ ²³⁸ U)	(²³⁰ Th/ ²³² Th)	(²³⁸ U/ ²³² Th)	[²³² Th] µg/g	[²³⁸ U] µg/g
Rock Standards									
BZVV (Secular Equilibrium Opal, n=4)	11	5	4	0.998 ± 0.006	0.989 ± 0.007	-	-	-	-
TML (Table Mountain Latite, n=2)	4	4	2	1.000 ± 0.002	0.991 ± 0.009	1.074 ± 0.011	1.084 ± 0.012	30.43 ± 0.72	10.79 ± 0.26
BCR-2 (USGS Rock Reference Material, n=3)	3	3	3	1.002 ± 0.003	1.001 ± 0.083	0.855 ± 0.057	0.854 ± 0.041	5.82 ± 0.68	1.71 ± 0.05
				Unweighted Average ²³⁴ U/ ²³⁸ U x 10 ⁻⁵	Error-Weighted Average* ²³⁴ U/ ²³⁸ U x 10 ⁻⁵	MSWD	Literature/Certified Value ²³⁴ U/ ²³⁸ U x 10 ⁻⁵	Reference	
Isotopic Standards									
CRM 145 (U isotope standard)	86			5.2841 ± 0.0188	5.2849 ± 0.0018	0.658	5.2841 ± 0.0082	New Brunswick Laboratory Certificate Value	
SRM 4321a (U activity standard)	9			5.2814 ± 0.0185	5.2810 ± 0.0078	2.627	5.2880 ± 0.0170	Oster et al. (2012), n=3	
				²³⁰ Th/ ²³² Th x 10 ⁻⁶	²³⁰ Th/ ²³² Th x 10 ⁻⁶		²³⁰ Th/ ²³² Th x 10 ⁻⁶		
IRMM-036 (Th isotope standard)		32		3.057 ± 0.028	3.059 ± 0.007	0.446	3.047 ± 0.040	Sims et al. (2008)	
OU Th "U" (Th isotope standard)		5		6.163 ± 0.069	6.157 ± 0.030	0.976	6.187 ± 0.075	Sims et al. (2008)	

* Calculated using ISOPLOT 3.75 (Ludwig, 2012)

TABLE A3. ²³⁰ Th-U GEOCHRONOLOGY MEASUREMENTS AND UNCORRECTED AGES. ALL UNCERTAINTIES REPORTED AT 2σ.															
Measured Concentrations			Measured Activity Ratios					Uncorrected Age			Single Sample Corrected Age				
Sample name	Coeval Sub-Sample ID	²³⁸ U (ppb)	²³² Th (ppb)	(²³⁰ U/ ²³⁸ U)	(²³⁰ Th/ ²³² Th)	(²³⁰ U/ ²³⁵ Th)	(²³⁰ Th/ ²³⁵ Th)	(²³⁰ Th/ ²³⁸ U)	Uncorrected Age (ka)	(²³⁰ Th/ ²³⁸ U) _{meas}	(²³⁰ Th/ ²³⁸ U) _{meas}	Corrected Age (ka)	(²³⁰ Th/ ²³⁸ U) _{meas}	Detrital Th Correction	
Accommodation Zone Shoreline Set															
SVD111-T2															
Inner Rind	T2-1a Inner	727.7 ± 2.0	281.3 ± 3.2	1.574 ± 0.005	7.907 ± 0.093	12.443 ± 0.149	3.477 ± 0.036	0.440 ± 0.002	35.01 ± 0.19	1.633 ± 0.005	-0.60	0.285 ± 0.022	21.49 ± 1.80	1.610 ± 0.006	0.52
	T2-1b Inner	604.4 ± 2.8	303.5 ± 2.9	1.565 ± 0.006	6.088 ± 0.065	9.529 ± 0.108	3.023 ± 0.032	0.497 ± 0.003	40.61 ± 0.32	1.634 ± 0.007	-0.57	0.295 ± 0.028	22.49 ± 2.36	1.602 ± 0.008	0.49
	T2-1c Inner	753.9 ± 1.9	327.4 ± 4.7	1.573 ± 0.010	7.037 ± 0.103	11.072 ± 0.176	3.147 ± 0.047	0.447 ± 0.002	35.70 ± 0.31	1.634 ± 0.011	-0.86	0.273 ± 0.025	20.53 ± 2.01	1.608 ± 0.011	0.25
	T2-1d Inner	750.1 ± 1.6	292.8 ± 2.9	1.571 ± 0.003	7.831 ± 0.079	12.302 ± 0.124	3.485 ± 0.041	0.445 ± 0.001	35.56 ± 0.16	1.631 ± 0.003	-0.44	0.288 ± 0.022	21.84 ± 1.82	1.607 ± 0.004	0.69
	T2-1e Inner	144.7 ± 0.5	100.7 ± 1.1	1.581 ± 0.003	4.393 ± 0.051	6.943 ± 0.082	2.482 ± 0.026	0.565 ± 0.003	46.85 ± 0.28	1.663 ± 0.004	-0.30	0.286 ± 0.039	21.48 ± 3.20	1.617 ± 0.007	0.85
Outer Rind	T2-2a Outer	548.2 ± 2.3	337.9 ± 4.2	1.577 ± 0.007	4.958 ± 0.066	7.820 ± 0.109	2.541 ± 0.030	0.512 ± 0.003	41.77 ± 0.34	1.649 ± 0.008	-0.65	0.245 ± 0.039	18.22 ± 3.10	1.608 ± 0.009	0.56
	T2-2b Outer	629.9 ± 1.5	402.4 ± 6.1	1.572 ± 0.006	4.784 ± 0.073	7.522 ± 0.118	2.455 ± 0.042	0.513 ± 0.002	42.00 ± 0.26	1.644 ± 0.006	-0.71	0.236 ± 0.040	17.56 ± 3.21	1.601 ± 0.008	0.66
	T2-2c Outer [†]	620.2 ± 2.5	521.6 ± 6.5	1.567 ± 0.013	3.634 ± 0.048	5.695 ± 0.088	2.640 ± 0.086	0.726 ± 0.003	65.04 ± 0.85	1.681 ± 0.014	-0.85	0.361 ± 0.053	28.14 ± 4.63	1.614 ± 0.016	0.46
	T2-2d Outer	609.5 ± 1.4	383.9 ± 3.6	1.587 ± 0.004	4.853 ± 0.048	7.703 ± 0.077	2.419 ± 0.028	0.498 ± 0.002	40.10 ± 0.20	1.658 ± 0.004	-0.54	0.225 ± 0.039	16.52 ± 3.10	1.615 ± 0.007	0.79
	T2-2e Outer	598.2 ± 1.8	422.3 ± 5.9	1.587 ± 0.009	4.330 ± 0.062	6.869 ± 0.106	2.329 ± 0.035	0.538 ± 0.002	43.93 ± 0.39	1.664 ± 0.010	-0.82	0.231 ± 0.044	17.04 ± 3.50	1.615 ± 0.012	0.50
SVD111-T3	T3-1a	627.3 ± 1.5	315.0 ± 3.7	1.509 ± 0.006	6.086 ± 0.073	9.183 ± 0.115	2.965 ± 0.030	0.487 ± 0.002	41.55 ± 0.28	1.572 ± 0.007	-0.76	0.333 ± 0.028	26.83 ± 2.53	1.549 ± 0.008	0.46
	T3-2a	631.7 ± 2.5	344.2 ± 4.1	1.519 ± 0.006	5.610 ± 0.071	8.519 ± 0.112	2.823 ± 0.031	0.503 ± 0.002	42.84 ± 0.32	1.585 ± 0.006	-0.60	0.336 ± 0.030	26.89 ± 2.73	1.560 ± 0.008	0.54
	T3-2b [†]	640.5 ± 4.0	316.5 ± 3.1	1.524 ± 0.003	6.185 ± 0.072	9.427 ± 0.110	2.859 ± 0.032	0.462 ± 0.003	38.55 ± 0.32	1.585 ± 0.003	-0.16	0.311 ± 0.028	24.52 ± 2.42	1.562 ± 0.005	0.74
	T3-2c [†]	672.2 ± 2.9	239.3 ± 2.5	1.520 ± 0.004	8.585 ± 0.098	13.046 ± 0.151	3.655 ± 0.043	0.426 ± 0.002	35.15 ± 0.23	1.574 ± 0.004	-0.38	0.316 ± 0.020	25.13 ± 1.77	1.558 ± 0.005	0.55
	T3-2d	655.4 ± 2.8	433.1 ± 5.5	1.513 ± 0.004	4.625 ± 0.062	6.999 ± 0.095	2.521 ± 0.036	0.545 ± 0.003	47.38 ± 0.32	1.587 ± 0.004	-0.33	0.342 ± 0.037	27.55 ± 3.33	1.555 ± 0.006	0.80
	T3-2e	660.9 ± 4.2	200.5 ± 2.1	1.523 ± 0.003	10.073 ± 0.125	15.342 ± 0.193	4.351 ± 0.049	0.432 ± 0.003	35.65 ± 0.30	1.578 ± 0.004	-0.20	0.339 ± 0.017	27.04 ± 1.56	1.565 ± 0.004	0.53
	T3-3a [†]	594.6 ± 1.3	381.0 ± 15.7	1.461 ± 0.010	4.770 ± 0.197	6.968 ± 0.292	2.392 ± 0.085	0.502 ± 0.002	44.79 ± 0.43	1.523 ± 0.011	-0.90	0.305 ± 0.038	25.22 ± 3.46	1.495 ± 0.012	0.36
SVD111-T4	T4-1a [†]	832.4 ± 3.9	539.8 ± 5.5	1.491 ± 0.046	4.713 ± 0.053	7.027 ± 0.229	2.342 ± 0.022	0.497 ± 0.003	43.18 ± 1.67	1.555 ± 0.049	-0.98	0.215 ± 0.050	16.87 ± 4.27	1.515 ± 0.047	0.00
	T4-1b-1	810.0 ± 3.8	407.7 ± 4.2	1.502 ± 0.017	6.072 ± 0.069	9.122 ± 0.148	2.886 ± 0.027	0.475 ± 0.003	40.55 ± 0.63	1.563 ± 0.019	-0.90	0.257 ± 0.039	20.22 ± 3.37	1.532 ± 0.019	0.19
	T4-1b-2 [†]	1080.5 ± 2.5	636.2 ± 7.3	1.526 ± 0.005	5.191 ± 0.061	7.920 ± 0.096	2.507 ± 0.040	0.483 ± 0.002	40.55 ± 0.24	1.590 ± 0.006	-0.69	0.227 ± 0.046	17.43 ± 3.77	1.552 ± 0.008	0.73
	T4-1b-3	635.0 ± 2.5	422.1 ± 4.3	1.492 ± 0.003	4.597 ± 0.051	6.859 ± 0.076	2.511 ± 0.036	0.546 ± 0.003	48.37 ± 0.30	1.564 ± 0.003	-0.30	0.257 ± 0.052	20.44 ± 4.47	1.521 ± 0.007	0.89
	T4-1b-4	655.7 ± 1.4	312.3 ± 4.0	1.525 ± 0.003	6.418 ± 0.083	9.785 ± 0.128	2.933 ± 0.043	0.457 ± 0.002	38.03 ± 0.17	1.584 ± 0.003	-0.44	0.250 ± 0.037	19.35 ± 3.11	1.554 ± 0.006	0.84
	T4-1b-5	658.8 ± 2.7	467.7 ± 6.6	1.483 ± 0.012	4.305 ± 0.064	6.387 ± 0.108	2.406 ± 0.039	0.559 ± 0.003	50.13 ± 0.61	1.557 ± 0.013	-0.86	0.251 ± 0.055	19.98 ± 4.79	1.511 ± 0.014	0.44
	T4-1c [†]	1078.0 ± 5.2	1057.5 ± 44.5	1.380 ± 0.042	3.115 ± 0.132	4.298 ± 0.224	1.686 ± 0.028	0.541 ± 0.003	52.93 ± 2.17	1.441 ± 0.046	-0.98	0.115 ± 0.076	9.46 ± 6.53	1.390 ± 0.044	0.12
	T4-1d	858.6 ± 8.9	781.8 ± 10.1	1.654 ± 0.003	3.357 ± 0.056	5.553 ± 0.092	1.694 ± 0.019	0.505 ± 0.005	38.71 ± 0.49	1.730 ± 0.003	0.13	0.110 ± 0.057	7.44 ± 3.99	1.668 ± 0.008	0.93
	T4-1e-2	590.6 ± 1.2	380.1 ± 4.9	1.728 ± 0.004	4.792 ± 0.062	8.282 ± 0.108	2.224 ± 0.024	0.464 ± 0.001	33.35 ± 0.15	1.800 ± 0.004	-0.48	0.187 ± 0.040	12.42 ± 2.78	1.754 ± 0.007	0.85
	T4-1a-3	576.4 ± 1.4	306.6 ± 7.4	1.727 ± 0.007	5.745 ± 0.139	9.921 ± 0.243	2.084 ± 0.098	0.363 ± 0.001	25.29 ± 0.16	1.781 ± 0.008	-0.76	0.132 ± 0.033	8.61 ± 2.26	1.745 ± 0.009	0.52
	T4-1a-4	578.7 ± 1.8	401.3 ± 4.7	1.707 ± 0.010	4.407 ± 0.054	7.522 ± 0.101	1.960 ± 0.025	0.445 ± 0.002	32.22 ± 0.26	1.774 ± 0.010	-0.81	0.144 ± 0.043	9.53 ± 2.99	1.726 ± 0.012	0.50
T4-1c-1	1218.3 ± 7.1	1207.0 ± 13.4	1.536 ± 0.006	3.085 ± 0.039	4.738 ± 0.062	1.887 ± 0.020	0.612 ± 0.004	53.63 ± 0.50	1.623 ± 0.006	-0.43	0.182 ± 0.062	13.65 ± 4.92	1.557 ± 0.010	0.78	
T4-1c-2	596.1 ± 2.9	521.2 ± 9.6	1.714 ± 0.006	3.495 ± 0.066	5.989 ± 0.115	1.928 ± 0.034	0.552 ± 0.003	41.17 ± 0.31	1.802 ± 0.006	-0.44	0.172 ± 0.055	11.46 ± 3.82	1.737 ± 0.010	0.81	
T4-1c-3	606.2 ± 3.0	376.2 ± 4.2	1.711 ± 0.006	4.924 ± 0.060	8.424 ± 0.106	2.128 ± 0.026	0.432 ± 0.002	31.10 ± 0.23	1.776 ± 0.006	-0.49	0.163 ± 0.039	10.83 ± 2.70	1.733 ± 0.008	0.66	
SVD111-T18	T18-1c-1	3091.3 ± 10.7	390.1 ± 5.5	1.223 ± 0.007	24.220 ± 0.354	29.631 ± 0.465	4.500 ± 0.054	0.186 ± 0.001	17.87 ± 0.14	1.235 ± 0.008	-0.81	0.131 ± 0.008	12.31 ± 0.81	1.231 ± 0.007	-0.03
	T18-1c-2	3152.8 ± 8.6	378.0 ± 4.6	1.237 ± 0.002	25.493 ± 0.319	31.542 ± 0.396	6.330 ± 0.076	0.248 ± 0.001	24.25 ± 0.11	1.254 ± 0.003	-0.46	0.196 ± 0.008	18.73 ± 0.82	1.250 ± 0.003	0.17
	T18-1d	3073.1 ± 16.9	321.3 ± 3.5	1.257 ± 0.022	29.231 ± 0.359	36.734 ± 0.786	3.393 ± 0.036	0.116 ± 0.001	10.53 ± 0.21	1.264 ± 0.023	-0.95	0.071 ± 0.007	6.30 ± 0.62	1.261 ± 0.022	-0.17
	T18-1e-1	2170.9 ± 7.4	359.9 ± 10.5	1.242 ± 0.023	18.435 ± 0.541	22.888 ± 0.793	3.496 ± 0.046	0.190 ± 0.001	17.98 ± 0.37	1.254 ± 0.024	-0.97	0.118 ± 0.011	10.83 ± 1.08	1.249 ± 0.023	-0.17
	T18-1e-2	3456.2 ± 8.4	346.9 ± 3.4	1.251 ± 0.003	30.447 ± 0.305	38.078 ± 0.386	7.825 ± 0.073	0.257 ± 0.001	24.89 ± 0.11	1.269 ± 0.003	-0.51	0.213 ± 0.007	20.28 ± 0.69	1.265 ± 0.003	0.11
	T18-1e-3	3595.4 ± 10.4	354.7 ± 5.0	1.252 ± 0.002	30.978 ± 0.447	38.799 ± 0.562	8.048 ± 0.116	0.260 ± 0.001	25.15 ± 0.12	1.271 ± 0.003	-0.44	0.217 ± 0.007	20.61 ± 0.73	1.268 ± 0.003	0.15
	T18-1e-4	3579.7 ± 7.1	354.6 ± 6.9	1.251 ± 0.003	30.849 ± 0.607	38.601 ± 0.763	8.051 ± 0.159	0.261 ± 0.001	25.31 ± 0.11	1.270 ± 0.003	-0.59	0.218 ± 0.008	20.74 ± 0.79	1.266 ± 0.003	0.12
	T18-1e-5	3579.7 ± 7.1	354.6 ± 6.9	1.251 ± 0.003	30.849 ± 0.607	38.601 ± 0.763	8.051 ± 0.159	0.261 ± 0.001	25.31 ± 0.11	1.270 ± 0.003	-0.59	0.218 ± 0.008	20.74 ± 0.79	1.266 ± 0.003	0.12
Middle Lake Shoreline Set															
SVD112-T1	T1a	726.0 ± 2.8	267.7 ± 2.6	1.499 ± 0.003	8.288 ± 0.088	12.420 ± 0.134	3.445 ± 0.034	0.416 ± 0.002	34.77 ± 0.21	1.550 ± 0.004	-0.36	0.256 ± 0.023	20.19 ± 2.00	1.528 ± 0.005	0.64
	T1b	738.4 ± 3.6	253.8 ± 2.5	1.483 ± 0.003	8.891 ± 0.097	13.185 ± 0.145	3.474 ± 0.036	0.391 ± 0.002	32.78 ± 0.22	1.530 ± 0.004	-0.30	0.242 ± 0.022	19.20 ± 1.87	1.510 ± 0.004	0.60
	T1c	734.2 ± 1.4	292.0 ± 3.6	1.476 ± 0.003	7.686 ± 0.096	11.342 ± 0.142	3.220 ± 0.038	0.419 ± 0.001	35.74 ± 0.16	1.526 ± 0.004	-0.55	0.246 ± 0.025	19.73 ± 2.18	1.503 ± 0.005	0.65
	T1d	761.6 ± 1.5	411.0 ± 14.7	1.454 ± 0.003	5.663 ± 0.203	8.235 ± 0.295	2.506 ± 0.034	0.443 ± 0.001	38.79 ± 0.18	1.507 ± 0.004	-0.56	0.208 ± 0.035	16.73 ± 2.98	1.476 ± 0.005	0.74
	T1e	760.2 ± 1.7	361.3 ± 3.4	1.461 ± 0.005	6.431 ± 0.062	9.397 ± 0.095	2.707 ± 0.026	0.421 ± 0.001	36.37 ± 0.20	1.511 ± 0.005	-0.66	0.215 ± 0.030	17.19 ± 2.56	1.484 ± 0.006	0.57
	T1f	764.8 ± 3.3	360.3 ± 4.1	1.462 ± 0.003	6.487 ± 0.079	9.487 ± 0.116	2.797 ± 0.041	0.431 ± 0.002	37.36 ± 0.24	1.514 ± 0.003	-0.33	0.227 ± 0.030	18.21 ± 2.57	1.487 ± 0.005	0.72
	T1g	782.3 ± 1.9	393.6 ± 7.2	1.457 ± 0.003	6.075 ± 0.112	8.853 ± 0.164	2.559 ± 0.051	0.421 ± 0.001	36.51 ± 0.17	1					

SVD112-T7	T7	372.5 ± 1.0	474.6 ± 7.0	1.518 ± 0.003	2.398 ± 0.036	3.642 ± 0.055	1.848 ± 0.064	0.771 ± 0.003	73.63 ± 0.43	1.638 ± 0.003	-0.31	0.218 ± 0.080	16.73 ± 6.57	1.543 ± 0.011	0.96	Basin Average (²³⁰ Th/ ²³² Th) _{basin}
SVD112-T9	T9-1a	582.8 ± 2.6	198.4 ± 2.0	1.832 ± 0.005	8.979 ± 0.101	16.453 ± 0.188	3.390 ± 0.033	0.378 ± 0.002	24.72 ± 0.16	1.893 ± 0.005	-0.43	0.194 ± 0.114	12.13 ± 7.49	1.861 ± 0.019	0.96	Rosholt Isochron (²³⁰ Th/ ²³² Th) _{basin}
	T9-1b	589.3 ± 3.0	198.2 ± 2.0	1.817 ± 0.007	9.087 ± 0.103	16.513 ± 0.194	3.351 ± 0.032	0.369 ± 0.002	24.32 ± 0.18	1.875 ± 0.007	-0.50	0.188 ± 0.113	11.81 ± 7.44	1.845 ± 0.019	0.93	
	T9-1c	586.7 ± 1.5	184.1 ± 1.9	1.818 ± 0.004	9.740 ± 0.104	17.708 ± 0.190	3.490 ± 0.036	0.358 ± 0.001	23.55 ± 0.11	1.874 ± 0.004	-0.45	0.190 ± 0.105	11.91 ± 6.94	1.846 ± 0.017	0.97	
	T9-1e	548.8 ± 1.1	182.3 ± 2.3	1.822 ± 0.011	9.202 ± 0.117	16.762 ± 0.234	3.415 ± 0.042	0.371 ± 0.001	24.42 ± 0.19	1.880 ± 0.011	-0.88	0.192 ± 0.112	12.08 ± 7.35	1.850 ± 0.021	0.84	
	T9-1f	593.5 ± 1.4	208.5 ± 2.8	1.815 ± 0.006	8.699 ± 0.117	15.787 ± 0.217	3.280 ± 0.045	0.377 ± 0.001	24.96 ± 0.13	1.874 ± 0.006	-0.64	0.188 ± 0.118	11.83 ± 7.79	1.842 ± 0.019	0.95	
	T9-1g ^a	596.6 ± 1.2	199.6 ± 2.4	1.811 ± 0.018	9.136 ± 0.111	16.548 ± 0.257	3.463 ± 0.053	0.379 ± 0.001	25.16 ± 0.29	1.871 ± 0.018	-0.95	0.199 ± 0.112	12.58 ± 7.47	1.841 ± 0.025	0.69	
	T9-1b+ ^a	629.4 ± 1.5	241.9 ± 3.0	1.768 ± 0.007	7.950 ± 0.101	14.056 ± 0.186	2.897 ± 0.047	0.364 ± 0.001	24.75 ± 0.15	1.824 ± 0.007	-0.74	0.158 ± 0.129	10.11 ± 8.63	1.790 ± 0.021	0.93	
SVD112-T10	T10a-1	586.7 ± 1.3	216.7 ± 2.6	1.808 ± 0.006	8.277 ± 0.102	14.969 ± 0.189	3.049 ± 0.035	0.368 ± 0.001	24.42 ± 0.13	1.866 ± 0.006	-0.67	0.273 ± 0.020	17.65 ± 1.41	1.850 ± 0.007	0.45	Rosholt Isochron (²³⁰ Th/ ²³² Th) _{basin}
	T10a-2	601.6 ± 1.6	189.7 ± 2.3	1.811 ± 0.007	9.690 ± 0.123	17.546 ± 0.231	3.375 ± 0.038	0.348 ± 0.001	22.93 ± 0.14	1.865 ± 0.007	-0.72	0.267 ± 0.017	17.20 ± 1.20	1.851 ± 0.008	0.31	
	T10a-3	589.3 ± 1.6	286.2 ± 3.0	1.768 ± 0.012	6.293 ± 0.069	11.125 ± 0.143	2.479 ± 0.023	0.394 ± 0.001	26.99 ± 0.24	1.829 ± 0.013	-0.88	0.269 ± 0.026	17.79 ± 1.88	1.807 ± 0.013	0.25	
	T10b-1 ^b	606.8 ± 2.5	159.7 ± 3.0	1.734 ± 0.008	11.615 ± 0.227	20.137 ± 0.352	3.850 ± 0.073	0.332 ± 0.002	22.80 ± 0.17	1.783 ± 0.008	-0.65	0.264 ± 0.015	17.80 ± 1.10	1.772 ± 0.008	0.22	
	T10b-2 ^b	594.0 ± 2.4	188.3 ± 2.1	1.827 ± 0.004	9.644 ± 0.195	17.618 ± 0.195	3.484 ± 0.045	0.361 ± 0.002	23.63 ± 0.14	1.884 ± 0.005	-0.40	0.280 ± 0.018	17.90 ± 1.21	1.870 ± 0.005	0.51	
	T10b-3	586.4 ± 1.1	199.2 ± 2.2	1.790 ± 0.004	8.996 ± 0.103	16.101 ± 0.250	3.162 ± 0.034	0.351 ± 0.001	23.46 ± 0.10	1.844 ± 0.005	-0.60	0.264 ± 0.019	17.21 ± 1.30	1.829 ± 0.006	0.52	
	T10b-4	1010.8 ± 2.0	263.3 ± 2.6	1.826 ± 0.004	11.733 ± 0.118	21.426 ± 0.218	3.958 ± 0.040	0.337 ± 0.001	21.93 ± 0.09	1.879 ± 0.004	-0.52	0.270 ± 0.014	17.26 ± 0.98	1.867 ± 0.005	0.47	
SVD112-T11	T11	1576.6 ± 4.2	631.3 ± 8.0	1.219 ± 0.003	7.632 ± 0.099	9.304 ± 0.122	3.418 ± 0.035	0.448 ± 0.002	49.18 ± 0.26	1.252 ± 0.003	-0.48	0.274 ± 0.025	27.56 ± 2.87	1.237 ± 0.003	0.55	Basin Average (²³⁰ Th/ ²³² Th) _{basin}
SVD112-T12	T12-1	648.7 ± 2.2	588.1 ± 7.2	1.243 ± 0.009	3.371 ± 0.043	4.189 ± 0.061	2.611 ± 0.027	0.775 ± 0.003	101.98 ± 1.46	1.323 ± 0.011	-0.85	0.381 ± 0.057	39.47 ± 7.02	1.271 ± 0.011	0.44	Basin Average (²³⁰ Th/ ²³² Th) _{basin}
	T12-2 ^a	2571.8 ± 11.8	1432.7 ± 18.8	1.194 ± 0.010	5.486 ± 0.076	6.549 ± 0.105	1.688 ± 0.022	0.308 ± 0.002	32.23 ± 0.37	1.212 ± 0.010	-0.84	0.066 ± 0.035	6.19 ± 3.36	1.197 ± 0.010	0.17	
	T12-3	1171.5 ± 3.3	1571.0 ± 18.2	1.162 ± 0.008	2.279 ± 0.027	2.649 ± 0.037	0.847 ± 0.003	0.847 ± 0.003	135.54 ± 2.35	1.238 ± 0.011	-0.88	0.265 ± 0.084	28.07 ± 10.07	1.176 ± 0.010	0.46	
	T12-4	1529.5 ± 3.2	1482.4 ± 14.4	1.193 ± 0.004	3.153 ± 0.031	3.762 ± 0.039	2.368 ± 0.023	0.751 ± 0.002	104.35 ± 0.85	1.259 ± 0.005	-0.70	0.331 ± 0.061	35.06 ± 7.53	1.213 ± 0.006	0.70	
	T12-5	1080.0 ± 2.6	1696.9 ± 17.4	1.166 ± 0.008	1.945 ± 0.021	2.269 ± 0.028	1.870 ± 0.023	0.961 ± 0.003	175.49 ± 3.61	1.273 ± 0.011	-0.87	0.280 ± 0.098	29.77 ± 11.92	1.181 ± 0.010	0.56	
	T12-6	132.1 ± 0.4	134.4 ± 1.4	1.180 ± 0.002	3.005 ± 0.034	3.545 ± 0.040	1.971 ± 0.023	0.656 ± 0.003	86.22 ± 5.8	1.230 ± 0.003	-0.40	0.215 ± 0.064	21.80 ± 7.11	1.191 ± 0.005	0.83	
	T12-7	140.2 ± 0.4	145.8 ± 1.5	1.176 ± 0.003	2.939 ± 0.032	3.456 ± 0.038	1.939 ± 0.021	0.660 ± 0.002	87.44 ± 0.59	1.225 ± 0.003	-0.47	0.209 ± 0.065	21.20 ± 7.26	1.187 ± 0.005	0.80	
Lower Lake Shoreline Set																
SVD112-T13	T13-1	606.7 ± 1.1	270.6 ± 2.9	1.541 ± 0.004	6.853 ± 0.075	10.563 ± 0.117	3.097 ± 0.029	0.452 ± 0.001	37.04 ± 0.18	1.601 ± 0.004	-0.61	0.277 ± 0.035	21.37 ± 2.91	1.575 ± 0.006	0.74	Rosholt Isochron (²³⁰ Th/ ²³² Th) _{basin}
	T13-2	593.8 ± 1.9	278.4 ± 2.6	1.532 ± 0.004	6.519 ± 0.065	9.983 ± 0.102	2.968 ± 0.031	0.455 ± 0.002	37.66 ± 0.21	1.591 ± 0.004	-0.47	0.272 ± 0.036	21.05 ± 3.07	1.564 ± 0.006	0.76	
	T13-3	594.4 ± 1.9	285.0 ± 3.0	1.517 ± 0.004	6.374 ± 0.070	9.670 ± 0.108	2.935 ± 0.031	0.461 ± 0.002	38.61 ± 0.22	1.577 ± 0.004	-0.48	0.273 ± 0.037	21.36 ± 3.18	1.549 ± 0.006	0.76	
	T13-4	590.2 ± 1.4	269.9 ± 2.7	1.520 ± 0.003	6.682 ± 0.070	10.155 ± 0.107	3.013 ± 0.036	0.451 ± 0.002	37.59 ± 0.18	1.578 ± 0.003	-0.46	0.272 ± 0.035	21.24 ± 3.03	1.552 ± 0.006	0.82	
	T13-5	633.2 ± 1.6	245.9 ± 2.5	1.525 ± 0.003	7.870 ± 0.082	12.005 ± 0.126	3.343 ± 0.033	0.425 ± 0.002	34.91 ± 0.17	1.580 ± 0.003	-0.47	0.273 ± 0.030	21.23 ± 2.56	1.558 ± 0.005	0.75	
	T13-6	607.5 ± 2.5	255.4 ± 2.7	1.520 ± 0.003	7.270 ± 0.083	11.048 ± 0.127	3.214 ± 0.038	0.442 ± 0.002	36.73 ± 0.22	1.576 ± 0.003	-0.29	0.277 ± 0.033	21.73 ± 2.80	1.553 ± 0.005	0.81	
	T13-7	605.8 ± 2.5	350.9 ± 3.5	1.515 ± 0.003	5.277 ± 0.057	7.995 ± 0.087	2.649 ± 0.028	0.502 ± 0.002	42.83 ± 0.27	1.581 ± 0.003	-0.27	0.275 ± 0.045	21.60 ± 3.85	1.548 ± 0.007	0.88	
SVD112-T14	T14-1	518.3 ± 2.4	475.9 ± 4.8	1.727 ± 0.004	3.329 ± 0.037	5.747 ± 0.065	2.037 ± 0.020	0.612 ± 0.003	46.17 ± 0.33	1.828 ± 0.004	-0.26	0.278 ± 0.102	18.91 ± 7.48	1.766 ± 0.017	0.97	Rosholt Isochron (²³⁰ Th/ ²³² Th) _{basin}
	T14-2	521.5 ± 1.1	282.0 ± 3.3	1.736 ± 0.009	5.651 ± 0.068	9.809 ± 0.129	2.676 ± 0.032	0.474 ± 0.002	33.95 ± 0.25	1.810 ± 0.010	-0.85	0.277 ± 0.060	18.72 ± 4.38	1.776 ± 0.014	0.68	
	T14-3 ^a	562.3 ± 1.8	369.4 ± 4.3	1.750 ± 0.010	4.652 ± 0.056	8.141 ± 0.108	2.205 ± 0.027	0.474 ± 0.002	33.66 ± 0.28	1.825 ± 0.011	-0.81	0.235 ± 0.073	15.57 ± 5.14	1.784 ± 0.015	0.73	
	T14-4	481.0 ± 1.1	283.1 ± 2.9	1.758 ± 0.004	5.193 ± 0.055	9.131 ± 0.097	2.592 ± 0.027	0.499 ± 0.002	35.52 ± 0.16	1.838 ± 0.004	-0.44	0.285 ± 0.065	19.04 ± 4.71	1.800 ± 0.011	0.94	
	T14-5	527.8 ± 1.4	298.1 ± 2.9	1.753 ± 0.008	5.411 ± 0.054	9.486 ± 0.104	2.603 ± 0.025	0.481 ± 0.002	34.17 ± 0.24	1.829 ± 0.009	-0.77	0.276 ± 0.063	18.43 ± 4.52	1.793 ± 0.013	0.76	
	T14-6	461.3 ± 1.0	345.3 ± 14.5	1.742 ± 0.007	4.083 ± 0.172	7.112 ± 0.301	2.306 ± 0.057	0.565 ± 0.002	41.50 ± 0.26	1.834 ± 0.008	-0.76	0.292 ± 0.084	19.79 ± 6.15	1.785 ± 0.016	0.87	
	T14-7 ^b	506.0 ± 1.5	371.9 ± 6.0	1.720 ± 0.009	4.158 ± 0.069	7.153 ± 0.124	2.331 ± 0.038	0.560 ± 0.002	41.76 ± 0.34	1.810 ± 0.010	-0.79	0.293 ± 0.082	20.11 ± 6.08	1.762 ± 0.016	0.80	
Upper Lake Shoreline Set																
SVD112-T15	T15b-1 ^a	609.2 ± 1.8	401.7 ± 6.4	1.569 ± 0.007	4.636 ± 0.076	7.276 ± 0.122	2.621 ± 0.045	0.565 ± 0.002	47.31 ± 0.34	1.651 ± 0.007	-0.70	0.290 ± 0.029	21.98 ± 2.44	1.606 ± 0.008	0.48	Rosholt Isochron (²³⁰ Th/ ²³² Th) _{basin}
	T15b-2	552.8 ± 1.2	329.6 ± 3.1	1.551 ± 0.004	5.125 ± 0.050	7.950 ± 0.078	2.600 ± 0.026	0.507 ± 0.002	42.13 ± 0.21	1.621 ± 0.004	-0.53	0.258 ± 0.026	19.62 ± 2.17	1.582 ± 0.005	0.68	
	T15b-3	594.4 ± 1.2	359.2 ± 3.4	1.564 ± 0.004	5.057 ± 0.049	7.910 ± 0.078	2.612 ± 0.025	0.517 ± 0.002	42.60 ± 0.20	1.636 ± 0.004	-0.53	0.264 ± 0.027	19.92 ± 2.19	1.597 ± 0.005	0.69	
	T15b-4 ^a	598.5 ± 1.1	481.6 ± 4.6	1.558 ± 0.003	3.798 ± 0.038	5.916 ± 0.059	2.133 ± 0.023	0.562 ± 0.002	47.38 ± 0.22	1.638 ± 0.003	-0.46	0.225 ± 0.035	16.87 ± 2.84	1.585 ± 0.006	0.82	
	T15b-5	612.7 ± 1.2	392.2 ± 4.4	1.551 ± 0.003	4.774 ± 0.055	7.405 ± 0.086	2.514 ± 0.030	0.527 ± 0.002	44.05 ± 0.20	1.624 ± 0.003	-0.45	0.259 ± 0.028	19.69 ± 2.33	1.583 ± 0.005	0.77	
	T15b-6	619.0 ± 1.4	574.2 ± 5.7	1.542 ± 0.003	3.295 ± 0.034	5.081 ± 0.053	2.133 ± 0.023	0.647 ± 0.002	57.27 ± 0.29	1.637 ± 0.003	-0.39	0.259 ± 0.041	19.87 ± 3.40	1.573 ± 0.006	0.86	
	T15b-7	610.7 ± 1.4	506.2 ± 5.0	1.559 ± 0.003	3.687 ± 0.037	5.748 ± 0.059	2.245 ± 0.022	0.609 ± 0.002	52.26 ± 0.26	1.648 ± 0.003	-0.40	0.262 ± 0.037	19.86 ± 3.00	1.591 ± 0.006	0.84	
SVD112-T17	T17-1	5516.6 ± 14.0	375.6 ± 4.9	1.161 ± 0.002	44.889 ± 0.599	52.118 ± 0.699	26.449 ± 0.334	0.589 ± 0.002	75.69 ± 0.45	1.199 ± 0.003	-0.44	0.560 ± 0.009	70.43 ± 1.55	1.196 ± 0.003	0.17	Basin Average (²³⁰ Th/ ²³² Th) _{basin}
SVD112-T18	T18b-1 ^a	417.0 ± 0.8	613.6 ± 7.6	1.406 ± 0.004	2.077 ± 0.026	2.921 ± 0.038	4.730 ± 0.063	2.277 ± 0.007	0.00 ± 0.00	0.000 ± 0.000	0.00	1.639 ± 0.095	0.00 ± 0.00	0.000		

TABLE A4. ISOCHRON AND MEASURED DETERMINATIONS OF $(^{230}\text{Th}/^{232}\text{Th})_{\text{initial}}$ FOR PALEOLAKE SURPRISE

Sample name	Details	$(^{230}\text{Th}/^{232}\text{Th})_{\text{initial}}^{\dagger}$
Accommodation Zone Shoreline Set		
SVDI11-T2 Inner Rind	4 Point Isochron	1.227 ± 0.085
SVDI11-T3	4 Point Isochron	0.938 ± 0.085
SVDI11-T4	4 Point Isochron	1.328 ± 0.119
SVDI11-T14	7 Point Isochron	1.137 ± 0.615
Middle Lake Shoreline Set		
SVDI12-T1*	6 Point Isochron	0.531 ± 0.479
SVDI12-T2	7 Point Isochron + Residue	0.904 ± 0.096
SVDI12-T5	5 Point Isochron + Residue	0.975 ± 0.236
SVDI12-T9	5 Point Isochron	1.610 ± 0.567
SVDI12-T10	5 Point Isochron	0.787 ± 0.083
Lower Lake Shoreline Set		
SVDI12-T13	7 Point Isochron	1.198 ± 0.118
SVDI12-T14	5 Point Isochron	1.112 ± 0.170
Upper Lake Shoreline Set		
SVDI12-T15	5 Point Isochron	1.279 ± 0.067
Error-Weighted Basin Average[§]	Tufa Isochrons (n=11) and Modern Carbonate (n=1)	1.327 ± 0.096
Sample name	Details	$(^{230}\text{Th}/^{232}\text{Th})$
SVDI11-C1	Modern Playa Carbonate	1.383 ± 0.016

* Not included in error-weighted average or used for single sample correction (Table DR3)

[†] Derived from the Rosholt isochron $(^{230}\text{Th}/^{232}\text{Th})$ vs $(^{238}\text{U}/^{232}\text{Th})$ intercept

[§] Used for basin average single sample correction (Table 5). Calculated using ISOPLLOT 3.75

TABLE A5. STABLE ISOTOPE MEASUREMENTS OF MODERN SURPRISE VALLEY WATERS

Sample Location	Material Type	δD	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	Reference
		(‰ VSMOW)	(‰ VSMOW)	(‰ VPDB)	
Bidwell Creek	Creek	-104	-14	-	Ingraham and Taylor (1989)
Bidwell Creek	Creek	-107	-15.1	-	Ingraham and Taylor (1989)
Bidwell Creek	Creek	-104	-14.8	-	Ingraham and Taylor (1989)
Bidwell Creek	Creek	-103	-14.6	-	Ingraham and Taylor (1989)
Bidwell Creek	Creek	-105	-15	-	Ingraham and Taylor (1989)
Bidwell Creek	Creek	-104	-14.3	-	Ingraham and Taylor (1989)
Bidwell Creek	Creek	-105	-13.7	-	Ingraham and Taylor (1989)
Cedar Creek	Creek	-105	-15	-	Ingraham and Taylor (1989)
Cedar Creek	Creek	-100	-15	-	Ingraham and Taylor (1989)
Powley Creek	Creek	-105	-14.2	-8	Sladek et al. (2004)
Spring (location unknown)	Groundwater	-115	-14.6	-	Ingraham and Taylor (1989)
Windmil (location unknown)	Groundwater	-113	-15	-	Ingraham and Taylor (1989)
Well (location unknown)	Groundwater	-113	-15.2	-14.2	Sladek et al. (2004)
Irrigation well	Groundwater	-106	-14.5	-14.2	Sladek et al. (2004)
Irrigation well	Groundwater	-106	-14.4	-14.9	Sladek et al. (2004)
Cockrel housewell	Groundwater	-105	-14	-13.3	Sladek et al. (2004)
Irrigatoin well	Groundwater	-109	-14.6	-14.8	Sladek et al. (2004)
Dean's housewell	Groundwater	-102	-13.7	-11.2	Sladek et al. (2004)
Cold brown water well	Groundwater	-115	-14.9	-	Sladek et al. (2004)
Middle Alkali Lake	Lake Water	-70	-9.5	-	Ingraham and Taylor (1989)
Middle Alkali Lake	Lake Water	-65	-8.7	-	Ingraham and Taylor (1989)
Middle Alkali Lake	Lake Water	-59	-7.2	-	Ingraham and Taylor (1989)
Middle Alkali Lake	Lake Water	-8	3.4	-	Ingraham and Taylor (1989)
Middle Alkali Lake	Lake Water	24	8.5	-	Ingraham and Taylor (1989)
Middle Alkali Lake	Lake Water	10	4.9	-	Ingraham and Taylor (1989)
Cedar Pass	Snow	-117	-15.7	-	Ingraham and Taylor (1989)
Dismall Swamp	Snow	-107	-14.8	-	Ingraham and Taylor (1989)
Dismall Swamp #2	Snow	-102	-16.4	-	Ingraham and Taylor (1989)
Hays Canyon	Snow	-107	-15.4	-	Ingraham and Taylor (1989)
North Star	Snow	-106	-16.8	-	Ingraham and Taylor (1989)
Reservation Creek	Snow	-107	-14.6	-	Ingraham and Taylor (1989)
49°23 Mt.	Snow	-110	-16.7	-	Ingraham and Taylor (1989)
Blue Lake	Snow	-119	-15.8	-	Ingraham and Taylor (1989)
Eagle Peak	Snow	-114	-16.3	-	Ingraham and Taylor (1989)
Barber Creek	Snow	-111	-14.4	-	Ingraham and Taylor (1989)

TABLE A6. INTERPOLATED CLIMATE MODEL PRECIPITATION, EVAPORATION, RUNOFF AND RELATIVE HUMIDITY PREDICTIONS FOR SURPRISE VALLEY (120°W, 41.5°N) FOR PREINDUSTRIAL AND 21 ka (LGM)

PMIP3 Climate Model	# of Model Years Averagd Preindustrial/LGM	Preindustrial Precipitation (mm/yr)	LGM Precipitation (mm/yr)	Preindustrial Total Evaporation (mm/yr)	LGM Total Evaporation (mm/yr)	Δ Precipitation* (%)	Δ Evaporation* (%)	Δ P-ET* (%)	Preindustrial Total Runoff (mm/yr)	LGM Total Runoff (mm/yr)	Δ Runoff* (%)
Modern Meterology		412 to 741 [†]		1401 [‡]							
CNRM-CM5	50/50	857	741	468	366	-13.6	-21.8	-3.8	387.7	371.6	-4.2
COSMOS-ASO	400/600	510	498	427	382	-2.2	-10.5	40.7	-	-	-
FGOALS-g2	10/10	1172	1359	435	237	16.0	-45.5	52.2	738.8	1112.4	50.6
GISS-E2-R	25/25	955	926	599	373	-3.1	-37.7	55.1	314.4	308.3	-1.9
IPSL-CM5A-LR	1000/200	733	819	471	433	11.8	-8.0	47.4	478.3	157.6	-67.1
MIROC-ESM	99/100	1293	1379	772	442	6.7	-42.8	80.0	521.7	935.5	79.3
MPI-ESM-P	150/100	791	1189	606	467	50.3	-22.9	289.2	179.3	324.9	81.2
MRI-CGCM3	500/100	1153	986	637	403	-14.5	-36.8	13.1	518.6	564.7	8.9
NCAR-CCSM4	156/31	644	693	501	363	7.6	-27.5	131.2	140.1	322.8	130.3
Ensemble Avg		900.9	954.5	546.2	385.1	6.5	-28.2	78.3	409.9	512.2	34.6
1σ		263.5	304.5	115.3	66.6	19.5	13.6	87.9	197.4	338.1	62.6
						n=9	n=9	n=9			n=8
LGM isotope-mass balance calculation:						9.5	-36.4				

	Preindustrial Mean Annual Relative Humidity (%)	LGM Mean Annual Relative Humidity (%)	Δ Annual Relative Humidity* (%)	Δ Monthly Relative Humidity* (%)	Preindustrial MAT (°C)	LGM MAT (°C)	Δ MAT * (°C)	Winter (DJF) Δ Precipitation* (%)	Summer (JJA) Δ Evaporation* (%)	Winter (DJF) MAT * (°C)	Δ Summer (JJA) MAT * (°C)	Δ
Modern Meterology	57.5 [§]				8.9 to 9.4 [†]							
CNRM-CM5	71.5	73.0	2.2	1.6	4.8	0.9	-3.9	-9.7	-13.2	-5.2	-1.8	
COSMOS-ASO	-	-	-	-	6.9	0.6	-6.3	-20.7	-11.1	-7.0	-5.7	
FGOALS-g2	-	-	-	-	6.8	-3.0	-9.8	3.3	-36.8	-7.5	-12.9	
GISS-E2-R	70.8	73.2	3.4	2.4	6.5	-0.3	-6.8	13.3	-40.4	-5.4	-8.0	
IPSL-CM5A-LR	-	-	-	-	4.3	-2.6	-6.9	2.5	-28.8	-5.1	-8.2	
MIROC-ESM	81.2	83.0	2.3	1.9	10.6	1.4	-9.3	-4.9	-33.3	-9.8	-9.5	
MPI-ESM-P	-	-	-	-	5.5	-1.2	-6.7	16.2	-30.8	-6.4	-7.0	
MRI-CGCM3	76.6	74.5	-2.7	-2.1	5.0	-0.2	-5.2	-12.1	-7.1	-5.5	-4.1	
NCAR-CCSM4	67.3	77.6	15.4	10.4	7.2	-0.5	-7.7	17.5	12.2	-6.4	-8.8	
Ensemble Avg	73.5	76.3	4.1	2.8	6.4	-0.6	-7.0	0.6	-21.0	-6.5	-7.3	
1σ	5.5	4.2	6.7	4.6	1.9	1.5	1.8	13.5	17.3	1.5	3.2	
			n=5	n=5			n=9	n=9	n=9	n=9	n=9	

* The percent (%) change relative to modern, Δ Precipitation, Δ Evaporation, Δ P-ET, and Δ Runoff is calculated as the 21 ka model output minus the preindustrial model output. Δ Humidity and Δ MAT is the subtraction of the 21 ka and preindustrial model outputs.

[†] Average range of precipitation, humidity and temperature from the Lake City (1929-1960), Cedarville (1894-2012) and Fort Bidwell (1911-2011) meteorological stations archived in the Western Regional Climate Center database (<http://www.wrcc.dri.edu>)

[§] Estimated Kimberly-Penman evapotranspiration and measured humidity from AgriMet (<http://www.usbr.gov/>) for Cedarville, CA

Table A7. COMPILATION OF WESTERN UNITED STATES LAKE HIGHSTANDS AND STILLSTANDS

Lake Name	Approximate Latitude	Highstand, stillstand or wet event	¹⁴ C age (yr) ± 1σ	Median Age (ka cal BP)	Calibrated age range (ka cal BP) ± 2σ*	Non-radiocarbon derived ages	Source
Carp Lake	45.9	Pollen Wet Event				9 - 14	Whitlock and Bartlein (1997) and Whitlock et al. (2000), as in Lyle et al. (2012)
Little Lake	44.2	Pollen Wet Event				11 - 15	Grigg and Whitlock (1998) and Worona and Whitlock (1995) as in Lyle et al. (2012)
Fort Rock/Alkali Lake	43.5	Highstand				17.7 - 18.3	Freidel (1993) as in Lyle et al. (2012)
Fort Rock/Alkali Lake	43.5	Highstand				19.5 - 21.4	Freidel (1993) as in Lyle et al. (2012)
Chewaucan Basin	42.8	Highstand	17500 ± 300	20.840	20.092 - 21.549		Allison (1982), recalculated using IntCal09
Chewaucan Basin	42.8	Stillstand	11930 ± 90	13.780	13.569 - 14.000		Licciardi (2001), recalculated using IntCal09 by Munroe and Laabs (2012)
Lake Surprise	41.5	Stillstand				19.13 - 21.82	<i>This study</i> , range of LGM radiocarbon ages from lower group 1 samples
Lake Surprise	41.5	Highstand	12750 ± 50	15.127	14.871 - 15.583		<i>This study</i> , calculated using IntCal09
Lake Surprise	41.5	Stillstand	12600 ± 50	14.878	14.500 - 15.199		<i>This study</i> , calculated using IntCal09
Lake Surprise	41.5	Stillstand	12420 ± 50	14.486	14.129 - 14.967		<i>This study</i> , calculated using IntCal09
Lake Surprise	41.5	Stillstand	10790 ± 50	12.670	12.572 - 12.831		<i>This study</i> , calculated using IntCal09
Clover Lake	40.9	Highstand	16361 ± 94	19.506	19.296 - 19.613		Munroe and Laabs (2012), calculated using IntCal09
Clover Lake	40.9	Highstand	14228 ± 71	17.310	16.978 - 17.629		Munroe and Laabs (2012), calculated using IntCal09
Waring Lake	40.8	Highstand	13800 ± 50	16.901	16.730 - 17.063		From K. Adams in Garcia and Stokes (2006), recalculated using IntCal09 by Munroe and Laabs (2012)
Lake Bonneville	40.5	Hydrologic Maximum				15 - 18.5	Inferred from Oviatt et al. (1992), compilation in McGee et al. (2012) and Miller et al. (2013)
Lake Franklin	40.3	Stillstand	15750 ± 140	18.924	18.651 - 19.345		Munroe and Laabs (2013), calculated using IntCal09
Lake Franklin	40.3	Stillstand	14250 ± 50	17.334	16.987 - 17.659		Munroe and Laabs (2013), calculated using IntCal09
Lake Franklin	40.3	Highstand	13250 ± 55	16.231	15.530 - 16.709		Munroe and Laabs (2013), calculated using IntCal09
Lake Franklin	40.3	Highstand	13150 ± 55	15.992	15.282 - 16.528		Munroe and Laabs (2013), calculated using IntCal09
Lake Franklin	40.3	Highstand	13600 ± 45	16.764	16.558 - 16.944		Munroe and Laabs (2013), calculated using IntCal09
Lake Franklin	40.3	Stillstand	11550 ± 50	13.387	13.267 - 13.537		Munroe and Laabs (2013), calculated using IntCal09
Lake Franklin	40.3	Stillstand	10950 ± 50	12.813	12.646 - 12.974		Munroe and Laabs (2013), calculated using IntCal09
Diamond Valley	40.0	Stillstand	13500 ± 180	16.562	15.599 - 16.988		Tackman (1993), recalculated using IntCal09 by Munroe and Laabs (2012)
Diamond Valley	40.0	Highstand	22060 ± 210	26.540	26.270 - 26.810		Tackman (1993), recalculated using IntCal09
Lake Lahontan	39.9	Highstand	13070 ± 60	15.800	15.200 - 16.413		Adams and Wesnousky (1998), recalculated using IntCal09 by Munroe and Laabs (2012)
Newark Valley	39.5	Stillstand	14480 ± 40	17.630	17.251 - 17.897		Redwine (2003) and Kurth et al. (2011), recalculated using IntCal09
Newark Valley	39.5	Stillstand	14450 ± 60	17.587	17.215 - 17.882		Redwine (2003) and Kurth et al. (2011), recalculated using IntCal09
Newark Valley	39.5	Highstand	13780 ± 50	16.889	16.730 - 17.063		Redwine (2003) and Kurth et al. (2011), recalculated using IntCal09 by Munroe and Laabs (2012)
Newark Valley	39.5	Stillstand	13690 ± 40	16.829	16.669 - 16.996		Redwine (2003) and Kurth et al. (2011), recalculated using IntCal09
Newark Valley	39.5	Stillstand	13025 ± 40	15.682	15.163 - 16.330		Redwine (2003) and Kurth et al. (2011), recalculated using IntCal09
Newark Valley	39.5	Stillstand	12375 ± 40	14.408	14.084 - 14.905		Redwine (2003) and Kurth et al. (2011), recalculated using IntCal09
Jakes Lake	39.3	Highstand	13870 ± 50	16.944	16.775 - 17.143		Garcia and Stokes (2006), recalculated using IntCal09 by Munroe and Laabs (2012)
Jakes Lake	39.3	Stillstand	13510 ± 40	16.694	16.439 - 16.886		Garcia and Stokes (2006), recalculated using IntCal09
Jakes Lake	39.3	Stillstand	12440 ± 50	14.527	14.148 - 14.989		Garcia and Stokes (2006), recalculated using IntCal09
Spring Lake	39.2	Stillstand	12710 ± 90	15.054	14.548 - 15.605		Lillquist (1994), recalculated using IntCal09 by Munroe and Laabs (2012)
Carpenter Lake	39.0	Stillstand	14210 ± 100	17.294	16.936 - 17.649		Lillquist (1994), recalculated using IntCal09 by Munroe and Laabs (2012)
Railroad Lake	38.9	Stillstand	13300 ± 120	16.276	15.479 - 16.821		Lillquist (1994), recalculated using IntCal09 by Munroe and Laabs (2012)
Lake Columbus	38.1	Highstand	14860 ± 50	18.255	17.795 - 18.519		Kurth et al. (2011), recalculated using IntCal09
Lake Russell (Mono)	38.0	Highstand				20 - 22.5	Benson et al. (1998)
Lake Russell (Mono)	38.0	Highstand	13100 ± 50	15.869	15.231 - 16.443		Benson et al. (1998), interpolated by and recalculated using IntCal09 by Munroe and Laabs (2012)
Lake Owens	36.5	Highstand	13240 ± 34	16.220	15.547 - 16.676		Bacon et al. (2006) and Orme and Orme (1993), combined in Oxcal† and recalculated using IntCal09 by Munroe and Laabs (2012)
Lake Owens	36.5	Highstand				19.5 - 27	Bacon et al. (2006)
Lake Manly	36.0	Highstand				20 - 22.5	Anderson and Wells (2003)
Searles Lake	35.8	Highstand				16 - 17	Lin et al. (1998)
Searles Lake	35.8	Highstand				22 - 27	Lin et al. (1998)
Mojave (1)	35.4	Highstand				20 - 22	Wells et al. (2003)
Mojave (2)	35.4	Highstand				13.5 - -17	Wells et al. (2003)
Thompson	35.2	Highstand				21 - 26	Orme (2008)
Thompson	35.2	Highstand				12.6 - 16.2	Orme (2008)
Lake Estancia	34.6	Highstand	13420 ± 100	16.556	15.877 - 16.892		Allen and Anderson (2000), recalculated using IntCal09 by Munroe and Laabs (2012)
Lake Estancia	34.6	Highstand	15480 ± 60	18.695	18.564 - 18.849		Allen and Anderson (2000), recalculated using IntCal09 by Munroe and Laabs (2012)
Lake Estancia	34.6	Highstand	15880 ± 60	19.100	18.837 - 19.361		Allen and Anderson (2000), recalculated using IntCal09 by Munroe and Laabs (2012)
Lake Estancia	34.6	Highstand	16730 ± 170	19.884	19.476 - 20.271		Allen and Anderson (2000), recalculated using IntCal09 by Munroe and Laabs (2012)
Lake Cochise	31.8	Highstand	13561 ± 60	16.889	16.730 - 17.063		Waters (1989), combined in Oxcal† and recalculated using IntCal09 by Munroe and Laabs (2012)

* Calibrated using the Calib 6.0 program with IntCal09 (Stuiver and Reimer, 1993; Reimer et al., 2009) by Munroe and Laabs (2012) or *This Study*.

† Ages combined in Oxcal 4.1 (Ramsey, 2006) by Munroe and Laabs (2012)

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Appendix B

Detailed description of ^{230}Th -U geochronology methods carried out in the Stanford University ICP-MS/TIMS Facility

B1. Introduction

The purpose of this appendix is to supplement the documentation of the laboratory methods developed and used for ^{230}Th - ^{238}U geochronology work in the ICP-MS/TIMS Facility at Stanford University. In Chapter 1 (Section 3.3) the analytical procedures used in this study have been documented for publication. This appendix is intended to provide additional details for future U-series projects carried out within the ICP-MS/TIMS Facility at Stanford University. The tables and figures presented herein represent the results of the initial method developed by multiple U-series users of our laboratory under the supervision of the facility director, Professor Kate Maher, and the laboratory managers, Dr. Caroline Harris (ICP-MS) and Dr. Karrie Weaver (clean laboratory). The facility has two mass spectrometers, a Nu Instruments Plasma HR multi-collector inductively coupled plasma mass spectrometer (MC-ICP-MS) and a Nu Instruments Attom high-resolution single-collector ICP-MS. Sample preparation is carried out in the PicoTrace metal-free clean laboratory with Class <1000 clean rooms.

B2. Sample dissolution and chromatographic methods

B2.1. Sample dissolution

Due to the varying solubility of Th and U (see discussion in Chapter 1) and potential for fractionation between U and Th we employed the “Total Sample Dissolution” method (e.g., Bischoff and Fitzpatrick, 1991; Luo and Ku, 1991). This method requires the complete dissolution of the carbonate samples, including any detritus (clays or organics) (cf. Ku, 2000). The sample dissolution procedure is briefly outlined in Section 3.3.3.1 of Chapter 1; the following table is a step-by-step description of sample dissolution for a typical carbonate sample containing both an insoluble and soluble fraction.

Table B1. Carbonate sample dissolution

Step	Details
Sample Cleaning	<ul style="list-style-type: none">- Sonicate carbonate chip (0.5 to 2 grams) in 18 MΩ deionized H₂O decant water then sonicate in ethanol. Move sample to new, clean glass beaker.- Dry overnight
Sample weighing	<ul style="list-style-type: none">- Weigh Teflon beaker (with cap)- Place sample in Teflon beaker and weigh- Calculate sample weight
Sample spiking	<ul style="list-style-type: none">- Add >3mL 18 MΩ deionized H₂O to cover sample- Spike sample based on a ²³⁶U/²³⁴U ratio of ~5 and sample weight
HNO ₃ dissolution	<ul style="list-style-type: none">- Carry out initial dissolution by adding concentrated HNO₃ (~14 M HNO₃)- Carefully add additional drops until CO₂ is not longer generated- Dry sample down at 120°C overnight- Treat sample residue with 5-10 drops of concentrated HNO₃- Dry sample to complete dryness at >120°C
H ₂ O ₂ treatment	<ul style="list-style-type: none">- Add a concentrated H₂O₂ – HNO₃ (1:10) mixture to cover sample- Heat gently at ~50°C and leave overnight capped- Dry sample to complete dryness at <100°C (usually 80°C)
Decant soluble	<ul style="list-style-type: none">- Bring sample up in 5mL of 3 M HNO₃- Decant soluble fraction only
Aqua regia treatment	<ul style="list-style-type: none">- Dry down insoluble fraction to complete dryness at ~120°C- Cover sample residue with a concentrated HNO₃ – HCl (1:3) mixture- Heat at 70°C for 1 hour loosely capped- Allow to sit cold for at least one day- Dry sample to complete dryness at >120°C
HF treatment	<ul style="list-style-type: none">- Add a concentrated HNO₃ – HF (5:1) mixture to cover sample- Flux at low temperature- Add additional HF if necessary- Dry to complete dryness at >120°C
Boric acid treatment	<ul style="list-style-type: none">- Treat sample with 0.5 mL concentrated HNO₃ and several drops of H₃BO₃ to drive off fluorides- Dry down and treat with concentrated HNO₃ if necessary
Prepare for column chemistry	<ul style="list-style-type: none">- Recombine the insoluble and soluble fractions.- Dry down and bring up in 10 mL 3M HNO₃- Flux at low temp (~50°C) for several hours or overnight- Load samples onto columns cold at room temperature

B2.2. Chromatographic methods

Column chemistry was carried out in the PicoTrace metal free laboratory in the ICP-MS/TIMS Facility at Stanford University. In an effort to improve the individual separation of U and Th from all other elements (Ca, Mg, Sr, Fe etc.) in the playa leachates, modern water samples, modern carbonates and spiked Pleistocene tufa samples, the column chemistry procedure was updated after the first few rounds of sample processing. The two methods using element specific cation exchange resins are outlined in **Tables B2** and **B3**. The reason for the switch was twofold. Firstly, sample loading for the Tru-Spec Resin in 1 to 4 mL 7 M HNO₃ (**Table B2**) is not time efficient (~ 6 to 12 hours) for large carbonate sample sizes (up to 2 grams dissolved). Secondly, column calibrations, using the Nu Attom ICP-MS, suggest that for carbonate samples the separation of U from these Th cuts is not sufficient. Unlike the carbonate samples, all water samples processed using the Tru-Spec Resin were time efficient (~3 hours). Results from column calibrations using the Nu Attom ICP-MS suggest that too much U ‘leaked’ into the prior Th collection, despite relatively high total U yields (>80 %) in the U collection.

Calibrated by Kimberly Lau and Dr. Karrie Weaver, in the ICP-MS/TIMS Facility, the UTEVA Resin column chemistry (modified from Potter et al. (2005) and Stirling et al. (2007)) is advantageous and efficient when processing large numbers of carbonate samples. The new recipe loads the samples in greater volume (5 to 10 L) and weaker acid (3 M HNO₃). This recipe also collects Th in stronger HCl (3M), which improves the separation of U and Th and increases the flow-rate of the column. Despite requiring an increased total volume that needs to pass through the Teflon columns, the UTEVA resin column chemistry recipe proved to be much more time efficient for carbonate samples (~3 to 5 hours total) and provided excellent apparent U and Th column yields, >95 % and >80 % respectively. Tufa sample procedural blanks were <3.3 pg of U and <15.3 pg of Th, and column blank blanks contributed <2.0 pg of U and <5.7 pg of Th.

In addition to the water, leachate, and carbonate samples, several reference materials were processed using the UTEVA resin column chemistry. Spiked aliquots of

natural rock standards TML (n=2), BCR-2 (n=3) and BZVV (n=4) were processed, yielding isotopic values and concentrations in agreement with previously published literature values (see also discussion in Section 3.3.3.4. of Chapter 1 and **Table A1**). The isotopic measurements of BZVV yielded average activity ratios of $(^{234}\text{U}/^{238}\text{U}) = 0.998 \pm 0.006$ and $(^{230}\text{Th}/^{238}\text{U}) = 0.989 \pm 0.007$ (n=4). The measured isotope ratios of TML were $(^{230}\text{Th}/^{232}\text{Th}) = 1.074 \pm 0.011$, $(^{234}\text{U}/^{238}\text{U}) = 1.000 \pm 0.002$ and $(^{230}\text{Th}/^{238}\text{U}) = 0.991 \pm 0.009$ with concentrations of $[\text{Th}] = 30.43 \pm 0.72$ ng/g and $[\text{U}] = 10.79 \pm 0.26$ ng/g. The measured isotope ratios of BCR-2 were $(^{230}\text{Th}/^{232}\text{Th}) = 0.855 \pm 0.057$, $(^{234}\text{U}/^{238}\text{U}) = 1.002 \pm 0.003$ and $(^{230}\text{Th}/^{238}\text{U}) = 1.001 \pm 0.083$ with concentrations of $[\text{Th}] = 5.82 \pm 0.68$ ng/g and $[\text{U}] = 1.71 \pm 0.05$ ng/g.

Despite these results, the apparent column yields for TML and BCR-2 were low (30 to 50 %) based on spiking with StUTh1. This low column yield may have been due to the high Fe content of these igneous rocks (relative to the carbonate or water samples). Thus, for future chromatography using this technique, a ‘clean-up’ cation exchange column is recommended before the element-specific UTEVA recipe.

Table B2. Column Chemistry Procedure using Tru-Spec Resin for separation of Th and U. This recipe was originally developed by Luo et al, (1997).

Step	Details	Eluted
Resin	~0.25mL TRU-Spec Resin 100-150 μ	
Cleaning	2x 1.25 mL 0.2 M HCl 2x 1.25 mL 0.1 M HCl - 0.3 M HF	
Conditioning	1 mL 7 M HNO ₃	
Loading Sample	1 to 4 mL 7 M HNO ₃	
Wash	4 x 1 mL HNO ₃ 2 x 1 mL 3 M HCl	Major Elements Rare Earth Elements
Collection	2 x 1 mL 0.2 M HCl 2 x 1mL 0.1 M HCl - 0.3 M HF 1 x 1mL 0.6 M HF	Thorium Uranium Uranium

Table B3. Column Chemistry Procedure using UTEVA Resin for separation of Th and U. This recipe was developed and calibrated at Stanford University by Kimberly Lau and Karrie Weaver modifying methods of Potter et al. (2005) and Stirling et al. (2007).

Step	Details	Eluted
Resin	~0.25mL UTEVA Resin 100-150 μ	
Cleaning	2x 2 mL 3 M HCl 2x 2 mL 0.1 M HCl - 0.3 M HF	
Conditioning	2 mL 7 M HNO ₃	
Loading Sample	5 to 10 mL 3 M HNO ₃	
Wash	3 x 1 mL HNO ₃	Major and Rare Earth Elements
Collection	1 x 1mL 3 M HHNO ₃ + trace HF 3 x 1 mL 3 M HCl 4 x 1mL 0.1 M HCl - 0.3 M HF	Thorium Uranium

B3. Isotopic analysis

B3.1. Analyses of standard reference materials

In addition to the reproducibility of the four standards discussed in Chapter 1 (**Figure B1**), Th Ames Metal, both spiked and un-spiked, was also analyzed repeatedly during the analytical sessions to assess long-term reproducibility. During the time period of the sample analyses and spike calibration the measured $^{230}\text{Th}/^{232}\text{Th}$ for Th Ames Metal was 0.0000005780 ± 179 ($n = 53$) (**Figure B2**). Th Ames Metal is a gravimetric standard rather than a certified isotope standard. The only published isotopic measurement is from Shen et al. (2002; $^{230}\text{Th}/^{232}\text{Th} = 0.0000006029 \pm 36$), but analyses from another Nu Instrument HR-MC-ICP-MS laboratory, at UC Davis, have made unpublished measurements of 0.000000588 to 0.0000000625 (personal communication, Cooper, K.M., 2013), at the high end of the range of values reported in this study. This difference may result from the methods used for correction of ^{232}Th tailing (onto ^{230}Th) and Ion Counter–Faraday gain correction. Alternatively, this discrepancy may indicate that Th Ames metal is an isotopically heterogeneous gravimetric standard among sub-samples distributed to isotope research groups.

Apart from the long-term reproducibility of standards (**Figure B1** and **B2A**), it is also important to demonstrate consistent and accurate measurements of Th and U isotopic values at varying instrumental sensitivities (beam intensity) and sample concentrations.

Because natural sample concentrations can be heterogeneous, this flexibility is important when measuring a suite of many ‘unknown’ samples on a given day. **Figures B2B and B3** demonstrate ability of the Nu MC-ICPMS to reproduce results for Th Ames Metal (n=53), IRMM-036 (n=32) and CRM-145 (n=86) at beam intensities from ~1 to ~9 volts on ^{232}Th and ^{238}U (measured on Faradays) with little change in total analytical error.

The relative contribution of instrumental standard error, bracketing standard drift, and the degree to which an isotopic reference material (used as a bracketing standard) is characterized, was assessed by evaluating all Th isotope measurements made as part of this study. **Figure B4A** demonstrates that at intensities of < 500 counts per second for ^{230}Th measured on collector IC0, the instrumental standard error for $^{230}\text{Th}/^{232}\text{Th}$ increases exponentially from ~0.25-0.5% to 0.5 to 4% (2σ) relative to analyses at higher ^{230}Th intensities. The bracketing standard, IRMM-035, is characterized by Sims et al. (2008) with an uncertainty of 0.89% (2σ), and typical plasma drift recorded by this bracketing standard is usually ~0.25% (2σ). Total analytical error (σ_{total}) is calculated as the square root of the sum of the three sources of error squared:

$$2\sigma_{total} = \sqrt{(2\sigma_{Instrument})^2 + (2\sigma_{Bracket})^2 + (2\sigma_{Std})^2} \quad (1)$$

Where $\sigma_{Instrument}$ is the mass spectrometer’s standard error for $^{230}\text{Th}/^{232}\text{Th}$, $\sigma_{Bracket}$ is the mass spectrometer’s drift delimited by the bracketing standard (IRMM-035) as measured before and after the unknown sample, and σ_{Std} is the uncertainty of the bracketing standard’s value. Thus, based on the total analytical error (**Figure B4B**) calculated via equation 1, it is not until less than ~250 counts on ^{230}Th that the total analytical precision for a given $^{230}\text{Th}/^{232}\text{Th}$ ratio is dominated by the instrumental standard error. At values greater than 250 cps, the total error is predominantly limited by the known isotopic composition of the bracketing standard, IRMM-035. Error was similarly calculated for all U-series measurements.

Figure B1. Long-term stability of uranium and thorium standard measurements at the Stanford ICP-MS/TIMS Facility using a Nu Instruments Plasma HR multi-collector inductively coupled plasma mass spectrometer (MC-ICP-MS). All error bars are 2σ . This figure is the same as **Figure A1**, see Appendix A for a complete figure caption.

Figure B2. (A) Long-term stability of Th Ames Metal measurements at the Stanford ICP-MS/TIMS Facility using a Nu Instruments Plasma HR multi-collector inductively coupled plasma mass spectrometer (MC-ICP-MS). As with CRM 145 and IRMM-036 samples were screened at the 2σ level (see Section 3.3.3.3, Chapter 1). (B) All of the Th Ames Metal measurements plotted against the ^{232}Th beam intensity (in Volts). Spiked and unspiked samples are distinguished by color. All error bars are 2σ per analysis.

Figure B3. All CRM 145 (n=86, (A)) and IRMM-036 (n=32, (B)) measurements plotted against ^{238}U and ^{232}Th beam intensity (in Volts). The long-term stability of these standards is shown in **Figure B1**. Due to the relative scarcity and small aliquots of OU Th ‘U’ and SRM 4321A obtained by the Stanford ICP-MS/TIMS Facility these samples were measured less frequently than IRMM-036 and CRM 145. All error bars are 2σ per analysis.

Figure B4. All $^{230}\text{Th}/^{232}\text{Th}$ measurements made as part of this study. **(A)** Standard error (% 2σ) vs. ^{230}Th beam intensity (counts per second (cps)). **(B)** Total analytical error (% 2σ) vs. ^{230}Th beam intensity (cps). The standard error is the Nu Instruments HR-MC-ICP-MS instrumental error only. Typically bracketing error measurements (on IRMM-035) is $\sim 0.25\%$ and the Sims et al. (2008) working group error is 0.89% . Thus, only at ^{230}Th intensity < 250 counts does the instrument's standard error typically dominate the final analytical $^{230}\text{Th}/^{232}\text{Th}$ error. See the Appendix B text for a discussion of how total analytical error is calculated.

B3.2. StUTh1 $^{229}\text{Th}/^{236}\text{U}$ spike calibration details

The calibration of the mixed spike, StUTh1, containing ^{229}Th and ^{236}U was carried out over the course of sample analysis. StUTh1 is a gravimetric dilution of a previously calibrated $^{229}\text{Th}/^{236}\text{U}$, known as Uc Spike, obtained from Dr. James Paces of the U.S. Geological Survey, Denver, Colorado. It is imperative that the spike calibration be accurate to ensure that the correct $^{230}\text{Th}/^{238}\text{U}$ ratios are calculated during data reduction. In order to calibrate StUTh1 for concentration two gravimetric normal solution were prepared and sequentially diluted. The CRM 145 β -dilution and Th Ames Metal β -dilution were spiked with StUTh1 and diluted to concentrations suitable for analysis on the Nu Instrument HR MC-ICP-MS. Three aliquots of CRM 145 and two aliquots of Th Ames Metal (labeled as γ -dilutions) were spiked and run before and over the course of the sample analysis period. The final calibrated concentrations can be found in **Table B4**. **Tables B5 and B6** list all of the measured $^{236}\text{U}/^{238}\text{U}$ and calculated ^{236}U concentration (ng/g). These measurements are plotted graphically in **Figure B5**. All dilutions and spiking was carried out gravimetrically and the 2σ errors include both the analytical and gravimetric errors.

The unweighted average concentrations of StUTh1 are $[^{236}\text{U}] = 0.96177 \pm 240$ ng/g (n=54) and $[^{229}\text{Th}] = 0.22317 \pm 196$ ng/g (n=58). This $^{229}\text{Th}/^{236}\text{U}$ ratio of 0.2392 ± 22 mol/mol is in agreement with the pre-dilution calibration provided by DR. Paces of 0.23952 ± 32 mol/mol. The total error associated with the unweighted concentrations of StUTh1 is 0.913 %. To reduce this error we also implemented an error-weighted mean using ISOPLOT 3.75 (Ludwig, 2012), which gives average concentrations of $[^{236}\text{U}] = 0.96188 \pm 36$ ng/g (n=54) and $[^{229}\text{Th}] = 0.22250 \pm 51$ ng/g (n=58). This $^{229}\text{Th}/^{236}\text{U}$ ratio is lower than the previous calibration, at 0.23840 ± 55 mol/mol. For all age calculations presented in this study we used the unweighted average concentrations of StUTh1.

Table B4. StUTh1 Calibrated ^{229}Th and ^{236}U concentrations

	^{229}Th (ng/g)	^{236}U (ng/g)	$^{229}\text{Th}/^{236}\text{U}$ (mol/mol)
Unweighted Average	0.22317 ± 0.00196 (0.878%)	0.96177 ± 0.00240 (0.249%)	0.23915 ± 0.00218 (0.913%)
Error-wighted Average¹	0.22250 ± 0.00051 (0.228%)	0.96188 ± 0.00038 (0.039%)	0.23840 ± 0.00055 (0.231%)
Previous Calibration²			0.23952 ± 0.00032 (0.135%)

¹ Calculated using ISOPLOT 3.75 (Ludwig, 2012)

² Provided by Dr. James Paces of the U.S. Geological Survey

Figure B5. StUTh1 ^{236}U and ^{229}Th concentration calibration using gravimetrically prepared spiked aliquots of CRM 145 ($n=3$) and Th Ames Metal ($n=2$). The solid black line is the unweighted mean concentration and the dashed black lines are the 2σ error.

Table B5. StUTh1 ^{229}Th Measurements

Analysis Date	$^{229}\text{Th}/^{232}\text{Th} \times 10^5$	^{229}Th Concentration (ng/g)	2 σ (%)	Th Ames Metal γ -dilution
May 10th 2012	2.20153 $\pm$ 0.03022	0.22353 $\pm$ 0.00307	1.37	γ 4 - dilution 1
May 10th 2012	2.19709 $\pm$ 0.03171	0.22308 $\pm$ 0.00322	1.44	γ 4 - dilution 1
May 10th 2012	2.19533 $\pm$ 0.02627	0.22290 $\pm$ 0.00267	1.20	γ 4 - dilution 1
May 20th 2012	2.19960 $\pm$ 0.03305	0.22333 $\pm$ 0.00336	1.50	γ 4 - dilution 1
May 20th 2012	2.20877 $\pm$ 0.03745	0.22426 $\pm$ 0.00380	1.70	γ 4 - dilution 2
May 20th 2012	2.21367 $\pm$ 0.03961	0.22476 $\pm$ 0.00402	1.79	γ 4 - dilution 2
May 20th 2012	2.21139 $\pm$ 0.03856	0.22453 $\pm$ 0.00392	1.74	γ 4 - dilution 2
May 20th 2012	2.19371 $\pm$ 0.02822	0.22274 $\pm$ 0.00287	1.29	γ 4 - dilution 2
May 20th 2012	2.21182 $\pm$ 0.03758	0.22457 $\pm$ 0.00382	1.70	γ 4 - dilution 2
May 20th 2012	2.20193 $\pm$ 0.03121	0.22357 $\pm$ 0.00317	1.42	γ 4 - dilution 2
May 20th 2012	2.19937 $\pm$ 0.03149	0.22331 $\pm$ 0.00320	1.43	γ 4 - dilution 2
May 20th 2012	2.20417 $\pm$ 0.03266	0.22380 $\pm$ 0.00332	1.48	γ 4 - dilution 2
May 20th 2012	2.21366 $\pm$ 0.03800	0.22476 $\pm$ 0.00386	1.72	γ 4 - dilution 2
May 20th 2012	2.20055 $\pm$ 0.04187	0.22343 $\pm$ 0.00425	1.90	γ 4 - dilution 2
May 20th 2012	2.19790 $\pm$ 0.03488	0.22316 $\pm$ 0.00354	1.59	γ 4 - dilution 2
May 20th 2012	2.21543 $\pm$ 0.03271	0.22494 $\pm$ 0.00332	1.48	γ 4 - dilution 2
May 20th 2012	2.21414 $\pm$ 0.04068	0.22481 $\pm$ 0.00413	1.84	γ 4 - dilution 2
May 20th 2012	2.19596 $\pm$ 0.02987	0.22296 $\pm$ 0.00303	1.36	γ 4 - dilution 2
April 5th 2013	2.20518 $\pm$ 0.02456	0.22390 $\pm$ 0.00249	1.11	γ 4 - dilution 2
April 5th 2013	2.19527 $\pm$ 0.02083	0.22289 $\pm$ 0.00212	0.95	γ 4 - dilution 2
May 17th 2013	2.19117 $\pm$ 0.01971	0.22248 $\pm$ 0.00200	0.90	γ 4 - dilution 2
April 5th 2013	2.40437 $\pm$ 0.01845	0.22242 $\pm$ 0.00171	0.77	γ 6
April 5th 2013	2.40321 $\pm$ 0.02140	0.22232 $\pm$ 0.00198	0.89	γ 6
April 5th 2013	2.40430 $\pm$ 0.02135	0.22242 $\pm$ 0.00198	0.89	γ 6
April 5th 2013	2.39864 $\pm$ 0.02155	0.22189 $\pm$ 0.00199	0.90	γ 6
April 5th 2013	2.40816 $\pm$ 0.02166	0.22278 $\pm$ 0.00200	0.90	γ 6
April 5th 2013	2.40324 $\pm$ 0.02151	0.22232 $\pm$ 0.00199	0.90	γ 6
April 5th 2013	2.40289 $\pm$ 0.02155	0.22229 $\pm$ 0.00199	0.90	γ 6
April 15th 2013	2.40263 $\pm$ 0.01777	0.22226 $\pm$ 0.00164	0.74	γ 6
April 15th 2013	2.41687 $\pm$ 0.01820	0.22358 $\pm$ 0.00168	0.75	γ 6
April 15th 2013	2.42238 $\pm$ 0.01972	0.22409 $\pm$ 0.00182	0.81	γ 6
April 15th 2013	2.42625 $\pm$ 0.01895	0.22445 $\pm$ 0.00175	0.78	γ 6
April 15th 2013	2.41317 $\pm$ 0.01852	0.22324 $\pm$ 0.00171	0.77	γ 6
April 15th 2013	2.42919 $\pm$ 0.01923	0.22472 $\pm$ 0.00178	0.79	γ 6
April 15th 2013	2.41556 $\pm$ 0.01925	0.22346 $\pm$ 0.00178	0.80	γ 6
April 18th 2013	2.40996 $\pm$ 0.03206	0.22294 $\pm$ 0.00297	1.33	γ 6
April 18th 2013	2.43858 $\pm$ 0.02981	0.22559 $\pm$ 0.00276	1.22	γ 6
April 18th 2013	2.41398 $\pm$ 0.02988	0.22331 $\pm$ 0.00276	1.24	γ 6
April 18th 2013	2.42334 $\pm$ 0.02886	0.22418 $\pm$ 0.00267	1.19	γ 6
April 20th 2013	2.41247 $\pm$ 0.03199	0.22317 $\pm$ 0.00296	1.33	γ 6
May 3rd 2013	2.40439 $\pm$ 0.01774	0.22243 $\pm$ 0.00164	0.74	γ 6
May 4th 2013	2.40026 $\pm$ 0.01801	0.22204 $\pm$ 0.00167	0.75	γ 6
May 4th 2013	2.41117 $\pm$ 0.01846	0.22305 $\pm$ 0.00171	0.77	γ 6
May 4th 2013	2.39048 $\pm$ 0.01847	0.22114 $\pm$ 0.00171	0.77	γ 6
May 17th 2013	2.39781 $\pm$ 0.02274	0.22182 $\pm$ 0.00210	0.95	γ 6 - dilution 1
May 17th 2013	2.41606 $\pm$ 0.02583	0.22351 $\pm$ 0.00239	1.07	γ 6 - dilution 1
May 17th 2013	2.40465 $\pm$ 0.02268	0.22245 $\pm$ 0.00210	0.94	γ 6 - dilution 1
May 17th 2013	2.40269 $\pm$ 0.02221	0.22227 $\pm$ 0.00205	0.92	γ 6 - dilution 1
May 17th 2013	2.40786 $\pm$ 0.02338	0.22275 $\pm$ 0.00216	0.97	γ 6 - dilution 1
May 17th 2013	2.40615 $\pm$ 0.02239	0.22259 $\pm$ 0.00207	0.93	γ 6 - dilution 1
May 17th 2013	2.42023 $\pm$ 0.02348	0.22389 $\pm$ 0.00217	0.97	γ 6 - dilution 1
May 17th 2013	2.41724 $\pm$ 0.02579	0.22362 $\pm$ 0.00239	1.07	γ 6 - dilution 1
May 17th 2013	2.40250 $\pm$ 0.01779	0.22225 $\pm$ 0.00165	0.74	γ 6 - dilution 1
May 17th 2013	2.40910 $\pm$ 0.01714	0.22286 $\pm$ 0.00159	0.71	γ 6 - dilution 1
May 17th 2013	2.40506 $\pm$ 0.01793	0.22249 $\pm$ 0.00166	0.75	γ 6 - dilution 1
May 17th 2013	2.39111 $\pm$ 0.01847	0.22120 $\pm$ 0.00171	0.77	γ 6 - dilution 1
May 17th 2013	2.39420 $\pm$ 0.01801	0.22148 $\pm$ 0.00167	0.75	γ 6 - dilution 1
May 17th 2013	2.41203 $\pm$ 0.01794	0.22313 $\pm$ 0.00166	0.74	γ 6 - dilution 1

Table B6. StUTh1 ^{236}U Measurements

Analysis Date	$^{236}\text{U}/^{238}\text{U} \times 10^5$	^{236}U Concentration (ng/g)	2 σ (%)	Th Ames Metal γ -dilution
April 4th 2013	6.23207 $\pm$ 0.01609	0.96109 $\pm$ 0.00248	0.26	γ 1
April 4th 2013	6.22596 $\pm$ 0.01720	0.96015 $\pm$ 0.00265	0.28	γ 1
April 4th 2013	6.23085 $\pm$ 0.01371	0.96090 $\pm$ 0.00212	0.22	γ 1
April 4th 2013	6.22818 $\pm$ 0.01408	0.96049 $\pm$ 0.00217	0.23	γ 1
April 4th 2013	6.22834 $\pm$ 0.01447	0.96052 $\pm$ 0.00223	0.23	γ 1
April 4th 2013	6.22625 $\pm$ 0.01441	0.96019 $\pm$ 0.00222	0.23	γ 1
April 4th 2013	6.24042 $\pm$ 0.01365	0.96238 $\pm$ 0.00211	0.22	γ 1
April 4th 2013	6.23239 $\pm$ 0.01365	0.96114 $\pm$ 0.00211	0.22	γ 1
April 4th 2013	6.24120 $\pm$ 0.01371	0.96250 $\pm$ 0.00211	0.22	γ 1
April 4th 2013	6.23751 $\pm$ 0.01446	0.96193 $\pm$ 0.00223	0.23	γ 1
April 4th 2013	6.23187 $\pm$ 0.01352	0.96106 $\pm$ 0.00208	0.22	γ 1
April 4th 2013	6.24101 $\pm$ 0.01412	0.96247 $\pm$ 0.00218	0.23	γ 1
April 4th 2013	6.22057 $\pm$ 0.01410	0.95932 $\pm$ 0.00217	0.23	γ 1
April 16th 2013	6.24354 $\pm$ 0.01787	0.96286 $\pm$ 0.00276	0.29	γ 1
April 16th 2013	6.23786 $\pm$ 0.01755	0.96198 $\pm$ 0.00271	0.28	γ 1
April 16th 2013	6.24625 $\pm$ 0.01807	0.96328 $\pm$ 0.00279	0.29	γ 1
April 16th 2013	6.23792 $\pm$ 0.02017	0.96199 $\pm$ 0.00311	0.32	γ 1
April 16th 2013	6.24412 $\pm$ 0.01966	0.96295 $\pm$ 0.00303	0.31	γ 1
April 18th 2013	6.24173 $\pm$ 0.01796	0.96258 $\pm$ 0.00277	0.29	γ 1
April 18th 2013	6.24628 $\pm$ 0.01779	0.96328 $\pm$ 0.00274	0.28	γ 1
April 18th 2013	6.24691 $\pm$ 0.01789	0.96338 $\pm$ 0.00276	0.29	γ 1
April 18th 2013	6.23701 $\pm$ 0.01738	0.96185 $\pm$ 0.00268	0.28	γ 1
April 16th 2013	6.17610 $\pm$ 0.01703	0.96227 $\pm$ 0.00265	0.28	γ 3
April 16th 2013	6.17115 $\pm$ 0.01590	0.96150 $\pm$ 0.00248	0.26	γ 3
April 16th 2013	6.16952 $\pm$ 0.01559	0.96125 $\pm$ 0.00243	0.25	γ 3
April 16th 2013	6.17253 $\pm$ 0.01664	0.96172 $\pm$ 0.00259	0.27	γ 3
April 25th 2013	6.17513 $\pm$ 0.01644	0.96212 $\pm$ 0.00256	0.27	γ 3
April 25th 2013	6.16644 $\pm$ 0.01626	0.96077 $\pm$ 0.00253	0.26	γ 3
April 25th 2013	6.16314 $\pm$ 0.01403	0.96025 $\pm$ 0.00219	0.23	γ 3
May 14th 2013	6.15369 $\pm$ 0.02238	0.95878 $\pm$ 0.00349	0.36	γ 3
May 14th 2013	6.15808 $\pm$ 0.02214	0.95946 $\pm$ 0.00345	0.36	γ 3
May 14th 2013	6.16559 $\pm$ 0.02232	0.96064 $\pm$ 0.00348	0.36	γ 3
May 14th 2013	6.16696 $\pm$ 0.02261	0.96085 $\pm$ 0.00352	0.37	γ 3
May 14th 2013	6.17516 $\pm$ 0.02262	0.96213 $\pm$ 0.00352	0.37	γ 3
May 14th 2013	6.16852 $\pm$ 0.02254	0.96109 $\pm$ 0.00351	0.37	γ 3
May 14th 2013	6.16724 $\pm$ 0.02257	0.96089 $\pm$ 0.00352	0.37	γ 3
May 14th 2013	6.17443 $\pm$ 0.02244	0.96201 $\pm$ 0.00350	0.36	γ 3
May 14th 2013	6.16533 $\pm$ 0.02237	0.96060 $\pm$ 0.00349	0.36	γ 3
May 14th 2013	6.17239 $\pm$ 0.02216	0.96170 $\pm$ 0.00345	0.36	γ 3
May 14th 2013	6.17513 $\pm$ 0.02249	0.96212 $\pm$ 0.00350	0.36	γ 3
May 14th 2013	6.17302 $\pm$ 0.02295	0.96179 $\pm$ 0.00358	0.37	γ 3
May 14th 2013	6.17942 $\pm$ 0.02246	0.96279 $\pm$ 0.00350	0.36	γ 3
April 16th 2013	6.18794 $\pm$ 0.01750	0.96423 $\pm$ 0.00273	0.28	γ 4
April 16th 2013	6.17937 $\pm$ 0.01741	0.96289 $\pm$ 0.00271	0.28	γ 4
April 16th 2013	6.17452 $\pm$ 0.01779	0.96214 $\pm$ 0.00277	0.29	γ 4
April 16th 2013	6.17422 $\pm$ 0.01761	0.96209 $\pm$ 0.00274	0.29	γ 4
April 16th 2013	6.17418 $\pm$ 0.01735	0.96209 $\pm$ 0.00270	0.28	γ 4
April 16th 2013	6.18562 $\pm$ 0.01773	0.96387 $\pm$ 0.00276	0.29	γ 4
April 16th 2013	6.16092 $\pm$ 0.01719	0.96002 $\pm$ 0.00268	0.28	γ 4
April 16th 2013	6.17833 $\pm$ 0.01892	0.96273 $\pm$ 0.00295	0.31	γ 4
May 14th 2013	6.18609 $\pm$ 0.01491	0.96394 $\pm$ 0.00232	0.24	γ 4
May 14th 2013	6.18109 $\pm$ 0.01359	0.96316 $\pm$ 0.00212	0.22	γ 4
May 14th 2013	6.17539 $\pm$ 0.01440	0.96227 $\pm$ 0.00224	0.23	γ 4
May 14th 2013	6.17398 $\pm$ 0.01424	0.96205 $\pm$ 0.00222	0.23	γ 4
May 14th 2013	6.18070 $\pm$ 0.01396	0.96310 $\pm$ 0.00218	0.23	γ 4

B4. Appendix B References Cited

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Appendix C

Processing Last Glacial Maximum climate model results

C1. Paleoclimate Model Intercomparison Project (PMIP) climate model ensemble

Comparison of inferred temperature and precipitation changes from paleoclimate records to climate models experiment provides a framework for model inter-comparison and validation. In this thesis the Surprise Valley results from the oxygen isotope model were compared to a climate model ensemble from PMIP3 (Braconnot et al., 2012; <http://pmip3.lsce.ipsl.fr/>). The nine available models are the same models used by the Intergovernmental Panel on Climate Change (IPPC) and are simulated for the LGM and the pre-industrial control (1950 AD). Thus, the forcings and boundary conditions are prescribed uniformly to simulate the LGM climate state. For the LGM greenhouse gas concentrations are lowered based on ice-core data (185 ppm for CO₂, 350 ppb for CH₄, and 200 ppb for N₂O), insolation changes are prescribed for 21 ka, and sea level is lowered by ~120 m globally to expand the size of the continents. Additionally, the presence of large ice-sheets, particularly in the Northern Hemisphere, is prescribed by modifying the surface albedo, ice sheet orography and ice sheet extent. Finally, the PMIP3 models, considered Earth System Models (ESM), contain dynamic simulation of vegetation change and the carbon cycle (Gent et al., 2011; Braconnot et al., 2012; Yukimoto et al., 2012; Brady et al., 2013; Voldoire et al., 2013; Kageyama et al., 2013a,b).

Due to computational limitations the PMIP3 models are simulated at varying horizontal and vertical resolutions, and for different lengths of time. **Table C3.1** lists the number of model years and the horizontal resolution. Bilinear interpolation of climate model data to a given location (see following section) is a common re-sampling technique that uses only the nearest four grid-points. By taking the product of the two linear functions that describe the horizontal gradients (in latitude-longitude space), a quadratic function calculates a spatially weighted value for the parameter of interest, such as precipitation, temperature or relative humidity.

Table C1. PMIP3 Models Processed. The horizontal resolution is dimensions of the atmospheric grid used by the model.

Model Name	Model Country	No. of Model Years (Preindustrial/LGM)	Horizontal Resolution
CNRM-CM5	France	50/50	256×128
COSMO-ASO	Germany	400/600	96×48
FGOALS-g2	China	10/10	128×60
GISS-E2-R	United States	25/25	144×90
IPSL-CM5A-LR	France	1000/200	96×95
MIROC-ESM	Japan	99/100	128×64
MPI-ESM-P	Germany	150/100	196×98
MRI-CGCM3	Japan	500/100	320×160
NCAR-CCSM4	United States	156/31	288×192

C2. Example NCAR Command Language (NCL) Scripts

The following two scripts, written in NCL, demonstrate how PMIP3 climate models are processed to produce monthly average values and interpolate parameters to a given location. This example uses the precipitation output from the climate model CNRM-CM5 (Voldoire et al., 2013) for the modern (pre-industrial) and LGM (~21 ka). The original model outputs are both monthly average precipitation for a fifty year climate model run (see **Table A6**). The first script takes the model output's netCDF file and uses the NCL function "clmMonTLL" to calculate the monthly average precipitation outputted as a new netCDF file for both the modern and LGM runs. These new twelve month netCDF climatology files were used to produce the maps in **Figures A13 and A14**. The second script manipulates the new netCDF files and interpolates to a given location using the NCL function "linint2_points_Wrap." This function uses bilinear interpolation (described previously) from the rectilinear grid of data. This example interpolates to Surprise Valley, CA (240°E, 41.5°N). Similar scripts were implemented for temperature, relative humidity, total evaporation, precipitation and runoff, for all available PMIP3 model outputs (n=9) to produce **Figures 8, 9, A11 and A12 and Table A6**.

C2.1. Climatology Script

```
load "$NCARG_ROOT/lib/ncarg/nclscripts/csm/gsn_code.ncl"
load "$NCARG_ROOT/lib/ncarg/nclscripts/csm/gsn_csm.ncl"
load "$NCARG_ROOT/lib/ncarg/nclscripts/csm/contributed.ncl"

.*****
;
; open file and read in monthly data from multi-year run

begin

  a=addfile("pr_Amon_CNRM-CM5_lgm_r1i1p1_195001-199912.nc", "r")
  LGM_pr=a->pr(:, :, :)
  b=addfile("pr_Amon_CNRM-CM5_piControl_r1i1p1_195001-199912.nc", "r")
  modern_pr=b->pr(:, :, :)

.*****
;
; use functions in contributed.ncl to produce monthly averaged variables

  LGM_P = clmMonTLL(LGM_pr)      ; monthly climatology(12,nlat,m lon)
  Modern_P = clmMonTLL(modern_pr) ; monthly climatology(12,nlat,m lon)

.*****
;
; Create netCDF files

  system("/bin/rm -f simple.nc") ; remove any pre-existing file
  ncdf1 = addfile("LGM_precip_CNRM.nc", "c") ; open output netCDF file
  ncdf2 = addfile("Modern_precip_CNRM.nc", "c") ; open output netCDF file

; make time and UNLIMITED dimensions
; recommended for most applications
  filedimdef(ncdf1, "time", -1, True)
  filedimdef(ncdf2, "time", -1, True)

; output variables directly into netCDF file
  ncdf1->LGM_P = LGM_P ; 3D
  ncdf2->Modern_P = Modern_P ; 3D

end
```

C2.2. Interpolation script

```
load "$NCARG_ROOT/lib/ncarg/nclscripts/csm/gsn_code.ncl"
load "$NCARG_ROOT/lib/ncarg/nclscripts/csm/gsn_csm.ncl"
load "$NCARG_ROOT/lib/ncarg/nclscripts/csm/contributed.ncl"
.*****
;
; open file and read in monthly data

begin

  precip_LGM = addfile("LGM_precip_CNRM.nc", "r")
  precLGM = precip_LGM->LGM_P(:, :, :)
  precip_Modern = addfile("Modern_precip_CNRM.nc", "r")
  precModern = precip_Modern->Modern_P(:, :, :)

; Define location for interpolation. These are the coordinates for Surprise Valley, CA
LON = (/ 240 /)
LAT = (/ 41.5 /)

; Define
lat_prec = precip_LGM->lat
lon_prec = precip_LGM->lon

.*****
;
; Interpolate precipitation values using linint2 function
CNRM_interp_prec_LGM = linint2_points_Wrap(lon_prec, lat_prec, precLGM, True, LON, LAT, 0)

CNRM_interp_prec_Modern = linint2_points_Wrap(lon_prec, lat_prec, precModern, True, LON,
LAT, 0)

.*****
;
; Display interpolated data
print(CNRM_interp_prec_LGM)
print(CNRM_interp_prec_Modern)

end
```

C3. Appendix C References Cited

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