AN ASSESSMENT OF LATE PLEISTOCENE TO MIDDLE HOLOCENE
LAKE LEVEL FLUCTUATIONS IN SURPRISE VALLEY,
CALIFORNIA

AN HONORS THESIS

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ABSTRACT

Knowledge of Earth's climate history and sensitivity, combined with modeling past and future climate, are central to informing policy decisions regarding future climate change. The hydrologic response to future warming scenarios due to increased anthropogenic CO₂ emissions remains uncertain. Freshwater availability in the arid western United States is projected to decrease in availability as increased agricultural, urban and industrial uses continue to stress supplies. Motivated by the potential for dramatic future hydrologic changes, studies recording the abrupt transitions between different equilibrium states of natural past climate variability shed light on our understanding of the modern climate system.

The presence of pluvial lakes in the Basin and Range Province, in the western United States, during the late Pleistocene (40 to 10 ka) indicates far greater moisture availability during the Pleistocene glacials. This study investigates the timing and magnitude of the most recent pluvial lake cycle that filled Surprise Valley, California using geophysical, geochemical and geochronologic tools. Spanning 31.2 to 4.6 ka, this new lake level record places the highest lake level, at 180 meters above present day playa, at 13.9 ± 1.2 ka. This age appears to be nearly synchronous with highstands of Lake Lahontan to the south and the Chewaucan Basin to the north. Additionally, most of the Basin and Range lake highstands, including Lake Surprise, follow peaks in precipitation minus evapotranspiration (P-ET) by 8-10 kyr. By compiling a diverse set of paleoclimate data available for western North America, I found that the timing and geographic distribution of lake highstands is inconsistent with increased precipitation in response to shifting westerly winds, the current model for the genesis of large lakes in western North America. Rather, lakes levels are more strongly correlated with changes in summer insolation, suggesting that lake highstands were likely facilitated by colder temperatures and increased humidity due to the presence of continental ice sheets and increased atmospheric convergence. I compared the constraints from lake and soil-based records to Atmosphere-Ocean General Circulation Model simulations from the Paleoclimate Model Intercomparison Project 2. Based on model-proxy intercomparison, the Atmosphere-Ocean General Circulation Models, the same models used to also assess future climatic changes, poorly predict hydrologic quantities for the Last Glacial Maximum.

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1. INTRODUCTION

Modern climate observations indicate that CO₂-induced temperature increases are intensifying the global hydrologic cycle (Huntington, 2006; Barnett *et al.*, 2008). Projecting into the 21st century, increased anthropogenic greenhouse emissions, primarily CO₂ from fossil fuel combustion, will cause changes in the seasonal and long-term availability of water, shifting the balance between precipitation and evapotranspiration (Seager *et al.*, 2007). Changes in the spatial patterns and amount of rainfall caused by this hydrologic intensification in the climate system will also affect water resources (Meehl *et al.*, 1999; Huntington, 2006), annual snow packs (Mote *et al.*, 2004; Barnett *et al.*, 2005), species migration (Walther *et al.*, 2002), and may result in more extreme weather events such as flooding and drought (Meehl *et al.*, 1999). Atmosphere-Ocean General Circulation Models (AOGCMs) are commonly used to model the effects of increased global temperatures, forced by increased anthropogenic CO₂ emissions. However, among AOGCM simulations there is widespread disagreement regarding future shifts in hydrologic parameters, both in the direction of change (e.g. increased or decreased precipitation) and the magnitude of change (Seager *et al.*, 2007; Solomon *et al.*, 2007). At the same time, accurately predicting changes in groundwater and surface water availability for agricultural, urban and industrial uses will become increasingly important.

The predictive power of AOGCM models can be assessed using historical climate observations as baselines. Recent evaluations of an ensemble of AOGCMs using paleoclimatic data indicate that models appear to reproduce the directional changes in modeled parameters (e.g. rainfall and evaporation) and the large-scale climate patterns found in the geologic record, but appear to underestimate the magnitude of regional changes (Braconnot *et al.*, 2012).

Reconstructing past changes in moisture availability during glacial-interglacial cycles places constraints on future changes in the hydrologic cycle in response to rising atmospheric CO₂. The same AOGCMs used to predict future climate scenarios (Solomon *et al.*, 2007) are also used to assess past climates (e.g. Braconnot *et al.*, 2007; Laîné *et al.*, 2009). The geologic record can be used to test the ability of AOGCMs to reproduce hydrologic and environmental responses to climatic forcings categorically different from those found within the modern or historical (late 1800s to present) climate record. Climate models simulating the Last Glacial Maximum (LGM, 19-26 ka) diverge in their prediction of past changes in rainfall amounts in the northeastern California/southern

Oregon region, which is situated in the transition zone between a more arid Pacific Northwest and a wetter central and Nevada during the LGM (Kim *et al.*, 2008; Laîné *et al.*, 2009). Proxy-model intercomparison can indicate geographic regions of large uncertainty, test the magnitude and direction of climate parameters calculated in the models, and determine the response of the climate system to feedbacks on local and global scales (Booth *et al.*, 2006; Braconnot *et al.*, 2012). These feedbacks may include sea-land-ice sheet dynamics, vegetation changes, and surface albedo effects.

Ultimately, the ability to make confident predictions using AOGCMs for future warming scenarios will require geographically extensive paleoclimatic data to test the models' ability to predict past climate. Thus, producing paleoclimate records with well-dated chronologies is important not just for our understanding of regional or global scale changes that occurred in the past, but also for future predictions of temperature, water availability, and seasonality. Paleoclimate data, such as those recorded in lake sediments and lake shorelines, can be used as a baseline for verification and interpretation of climate models (Booth *et al.*, 2006). Paleoclimate data reconstructions are abundant in the oceans and at high latitudes in ice cores, but there are limited records from the continents at mid-latitudes. Examples include $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data from cave deposits (e.g. Oster *et al.*, 2009; Wagner *et al.*, 2010), basin analysis from lake sediment cores (e.g., Negrini *et al.*, 2000; Cohen *et al.*, 2000; Licciardi *et al.*, 2001), dating of lake shoreline deposits (e.g. Oviatt *et al.*, 1992; Garcia and Stokes, 2006), dating fluvial terraces (e.g. Sharp *et al.*, 2003) and $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data from soil carbonates (e.g. Amundson *et al.*, 1996). This study seeks to fill gaps temporally and geographically in the paleoclimate record of the western United States. By constraining the timing and magnitude of fluctuations of the two lake systems in northeastern California (Surprise Valley) and southern Oregon (Chewaucan Basin), this study will increase the spatial coverage of paleoclimate data used to evaluate and test the different climate models in a region where they disagree.

2. BACKGROUND

2.1 Late Pleistocene Pluvial Lakes in Western North America

Vast pluvial lakes occupying inward draining basins covered the landscape of western North America during the late Pleistocene (Figure 1). The presence of large lakes in the now-arid valleys of the Basin and Range during Pleistocene glacial periods points to a fundamentally different hydrologic balance, requiring either increased precipitation and/or decreased evaporation during the late Pleistocene glacial periods (Benson *et al.*, 1990; Reheis, 1999a). Additionally, decreased temperature and increased relative humidity may have played a role in both increasing precipitation and/or decreasing evaporation. The net effect of the temperature changes is enhanced runoff within a given watershed, and diminished surface evaporation, leading to hydrologic conditions capable of creating and maintaining a lake system (Reheis, 1999a).

A lake's shoreline erosional features, along with the sedimentary facies, minerals and organic materials found in lake sediment cores, record the age of lake levels and the paleoenvironmental conditions (Adams and Wesnousky, 1998; Felton *et al.*, 2006; Godsey *et al.*, 2011). Therefore, the timing of fluctuations during past lake cycles can be determined by dating beach deposits such as tufa (Ku *et al.*, 1998; Garnet *et al.*, 2004) and studying lake sediment cores (e.g. Licciardi *et al.*, 2000; Zic *et al.*, 2002). Lake records provide spatially and temporally widespread paleoclimate data used to identify climatic processes driving precipitation patterns during the late Pleistocene. Small, closed basin pluvial lakes, such as those considered here, are perhaps most sensitive to changes in surface hydrology caused by increases in effective moisture (e.g. Licciardi, 2001; Garcia and Stokes, 2006). In the Great Basin, the chronology of lake highstands and later stillstands (e.g. periods of lake level equilibrium below the highstand) indicates a peak in moisture availability during and following the Last Glacial Maximum (LGM, 19-26ka) (Tchakerian and Lancaster, 2002) and also records the subsequent transition to more arid conditions during the Pleistocene to Holocene deglaciation. However, the atmospheric and hydrologic mechanisms explaining the late, post LGM lake highstands is still an outstanding research question in the paleoclimatology of western North America.

Paleoshoreline age records in pluvial lake basins also enhance ongoing research in the fields of paleohydrology, paleoenvironment and neotectonics. Distinct, constructional shorelines are excellent recorders of lake-level fluctuations because they are rapidly formed by wave action (Scholfield *et al.*, 2004; Felton *et al.*, 2006; Jewell, 2007; Jewell

and Nicoll, 2011). Shoreline age data constrain short-term changes in the hydrologic balance of a particular lake system. Additionally, as horizontal geomorphic features, shorelines are paleohorizontal datums that have been used to measure both brittle faulting and broad unwarping of the lithosphere (Bills and May, 1987; Adams and Wesnousky, 1998; Oldow and Singleton, 2008). However, due to subsequent erosion, isostatic rebound and eolian inputs, natural variability is often observed in shoreline deposits (Scholfield *et al.*, 2004; Jewell, 2007; Jewell and Nicoll, 2011) and it is important to recognize these processes when interpreting elevation data. Past studies have used a variety of techniques to quantify a lake's past hydrologic budget, climatic history and paleoenvironment. Interpreting results from geomorphic mapping, geophysical methods (such as terrestrial LiDAR on shoreline features), geochemistry (stable isotope analysis and elemental analysis of sediments or shoreline deposits), and geochronology (U-series, ^{14}C dating and/or cosmogenic nuclides) enable the geologic history of a lake system to be unraveled.

Figure 1. Map of the Western United States with Late Pleistocene Pluvial Lakes and Additional Paleoclimate Records (adapted from Reheis, 1999b). The yellow outline is the EPA defined Basin and Range province based on watershed regions. The dark blue areas indicate basins filled with lake systems during the late Pleistocene and light blue are modern lakes. The red areas are Lake Surprise and the Chewaucan Basin lake systems (this study). The additional paleoclimate records, used for comparison are found denoted by the black stars and circles. Data from previous studies includes a ΔP -ET proxy using $(^{234}\text{U}/^{238}\text{U})_0$ for pedogenic carbonates from Wind River, Wyoming (WR) (Sharp *et al.*, 2003), deep vadose zone opal from Yucca Mountain (YM) (Paces *et al.*, 2010) and pedogenic opal from Diamond Valley, Nevada (DV), Newark Valley, Nevada (NV), Fish Lake Valley, Nevada (FLV) and Barstow, California (BA) (Maher *et al.*, in prep); vein calcite from Devils Hole, Nevada (DH) (Winograd *et al.*, 2006); speleothem from Cave of the Bells, Arizona (COB) (Wagner *et al.*, 2010); and the timing of lake highstands at Lake Bonneville, Utah (Oviatt *et al.*, 1992); Lake Lahontan, Nevada (Adams and Wesnousky, 1998); Lake Newark, Nevada (Kurth *et al.*, 2011); Lake Diamond, Nevada (Tackman, 1993); Lake Jakes Lake, Nevada (Garcia and Stokes, 2006); Lake Manly, California (Ku *et al.*, 1998) and Searles Lake, California (Lin *et al.*, 1998).

2.2 Geologic and Climatic Setting

Extensional faulting in the Basin and Range province in the western United States formed numerous N-S-trending basins throughout the region. Most are closed basins (also referred to as terminal or endorheic basins), such that the basin retains all precipitation and allows no outflows of water to adjacent basins. Many of these closed basins sustained pluvial lakes during glacial periods of the Pleistocene but have since diminished in size or developed playa lakes during the Holocene (Figure 1) (Reheis, 1999a; Orme, 2008). Playas contain shallow lakes seasonally (during winter months) or during prolonged wet spells such as El Nino years (Orme, 2008).

In the late Pleistocene, highstands in these closed basins appear to postdate the LGM (Garcia and Stokes, 2006). However, changes in lake water levels are not necessarily synchronous between adjacent or nearby basins. These variations may be due to changes in the storm tracks that deliver moisture, basin-to-basin variability in water storage capacity, the surface area available to evaporation relative to lake volume, and linkages to adjacent basins (Orme, 2008). Lake level fluctuations in many systems (e.g. Bonneville, Lahonton, Russell, and Searles shown in Figure 1) have been attributed to the position of the mean winter westerly storm tracks (or jet stream) that bring winter precipitation to the region (Benson *et al.*, 1990; Benson *et al.*, 1998). The latest Pleistocene lake highstand and geochemical variations in the Chewaucan Basin have also been explained by changes in the position of the mean westerly storm tracks (Licciardi, 2001; Negrini, 2002; Zic *et al.*, 2002), but this record has not been compared to other basins. Other research on smaller closed basins, such as Lake Franklin, NV (Figure 1), has concluded that local lake/atmosphere feedback processes, including parameters such as increased cloud albedo, increased relative humidity and lowered net evaporation, better explain the presence of lake systems during glacial periods of the late Pleistocene (Garcia and Stokes, 2006).

Surprise Valley, CA, and the Chewaucan Basin, OR, on the western edge of the Basin and Range (Figure 1) capture precipitation brought by the westerly storm systems. Both basins are geographically positioned closer to the Pacific Ocean relative to other lake records from the Basin and Range. Thus, changes in lake levels should be more directly linked to changes in the intensity and geographic position of westerly storm tracks. The following is a summary of the geologic and climatic setting for the main study site, Surprise Valley, CA, and the Chewaucan Basin, OR.

Figure 2. Maps of the Surprise Valley Watershed and the Primary Study Area (Previous Page). (A) Shaded relief map of Surprise Valley (1:750:000) with the Environmental Protection Agency (EPA) Surprise Valley watershed outlined in magenta. The highstand elevation estimated by Reheis (1999b) is denoted by the red outline. The location of the three weather stations used in this study (data summarized in Figure. 3) is denoted by black diamonds. (B) Simplified geological map of the study area (1:60,000), with a Digital Elevation Model (DEM) and USGS 100k topography map used as a baselayer, adapted from Egger and Miller (2011) ArcGIS database and field observations, with observed and inferred faults (black lines), the Reheis (1999b) highstand elevation (magenta lines) and GPS elevation transects (brown lines, numbered T1-8 for the handheld GSP transects and T-USGS for the continuous transect obtained from Dr. Jonathan Glen). EPA's watershed regions are a publicly available dataset (http://cfpub.epa.gov/surf/huc.cfm?huc_code=18080001) and the stream data are based on high-resolution elevation data obtained from NASA's Shuttle Radar Topography Mission and made publicly available by the USGS HydroSHEDS program (<http://hydrosheds.cr.usgs.gov/>).

2.2.1 Surprise Valley, California

Surprise Valley (41.53°N, 120.17°W), in northeast California, sits on the northwest margin of the Basin and Range province (Figures 1 and 2). The N-S trending valley is a high altitude desert with three seasonal playa lakes. Surprise Valley is bounded on the west by the Surprise Valley fault, but the eastern margin is more complex, and the valley transitions from a full-graben in the south to a half-graben in the north (Personius *et al.*, 2008; Egger *et al.*, 2009; Lerch *et al.*, 2009; Egger and Miller, 2011). The accommodation zone between the half-graben in the north upper lake basin and the full graben in the middle and lower lake basins contains the most lake-related features, such as terraces and visible shorelines sets.

The modern watershed area demarcation by the Environmental Protection Agency (EPA) and USGS HydroSHEDS stream analysis demonstrate that Surprise Valley is a closed, terminal basin (Figure 2A). As a terminal basin, the establishment of a seasonal, decadal or millennial lake at Surprise Valley should be directly correlated to precipitation falling within the area of the watershed. The highstand elevation reported by Reheis (1999b) along with the modern topography suggests that minimal overflow occurred at the highstand to the south into Duck Flat. Thus, Duck Flat is considered part of the lake system.

It was recognized by early geologists that the highstand shoreline and sets of lower elevation shorelines appeared laterally continuous through the valley and into Duck Flat. The presence of an ancient lake system in Surprise Valley was first postulated by I.C. Russell (1884), and was first surveyed and described extensively by R.J. Russell (1927). Previous mapping and GPS transect work by Zimbelman *et al.* (2009), Sack (2002), Elder and de la Fuente (2009) and Reheis (1999b) suggests that the high stand, and thus the receding shorelines of Lake Surprise are similar to late Pleistocene (<100 ka) lakes commonly seen in other valleys in the Basin and Range province. Zimbelman *et al.* (2009) measured the high stand shoreline in the south part of the valley at 1533.6 meters whereas Reheis (1999b) reports 1567 meters (Figure 2A). Because Surprise Valley is thought to have been a closed basin with no overflow (Sack, 2002), highstand shorelines likely correspond to glacial periods as previously found in other Basin and Range terminal basins. Preliminary work by Elder and Juan de Fuca (2009) suggests that the highstand is Marine Isotope Stage (MIS) 2 (~17 ka) based on dating of landslide deposits that were cut and reworked by Pleistocene Lake Surprise in the north end of the valley.

The primary purpose of this study was to date both the highstand shoreline and lower stillstand shorelines using ^{230}Th -U geochronology of lake tufas found at the shoreline surfaces on the exposed bedrock.

The meteorological data shown in Figure 3 demonstrates why a modern lake does not persist year-round in Surprise Valley. Precipitation and relative humidity roughly covary, with increased humidity and precipitation occurring during the winter months. Predictably, temperature and evapotranspiration also covary, with increased temperatures and increased evapotranspiration occurring during the summer months. The net difference between precipitation and evapotranspiration (P-ET) of the historical meteorological data sets (see references in Figure 3 caption) indicate an average net loss of water from the basin of -6.96 cm/month, with a yearly P-ET of -0.83 cm. During a typical year, playa lakes form during the wetter, colder winter season in all three sub-basins (upper, middle, and lower) of Surprise Valley. During December to February, P-ET is positive, averaging 6.03 cm/month. Seasonal swings in the relative humidity in the historical meteorological data show an average relative humidity of 72.43 % during the winter months, whereas during summer months relative humidity averages 42.80 %. A summary of the hydrologic and climatic data for the Surprise Valley watershed can be found in the Appendix, Table A1.

Figure 3. Historical Hydrologic and Climate Data for the Surprise Valley Watershed. (A) Precipitation, evapotranspiration and calculated precipitation minus evapotranspiration (P-ET). Units are all m/month. (B) Mean, high and low temperature (°C), and relative humidity (%). The precipitation and temperature data is a composite average from three historical datasets maintained by the Western Regional Climate Center (WRCC, <http://www.wrcc.dri.edu/>) spanning 1894-2012 from the Cedarville, Lake City and Fort Bidwell, CA weather stations (all sites within Surprise Valley, see Figure 2B). Both temperature and precipitation are directly measured with precipitation being both rain and snow amounts, adjusted for liquid water amounts. The evaporation and relative humidity data was obtained from the Pacific Northwest Cooperative Agricultural Weather Network database (AgriMet, <http://www.usbr.gov/pn/agrimet/index.html>) for Cedarville, CA; the database spans the period 1985-2011. Relative humidity is directly measured and evaporation data is computed using the Kimberly Penman procedure (Dockter, 1994) using other climate variables such as wind speed, surface water availability, and temperature.

2.2.2 Lake Chewaucan Basin, Oregon

Summer Lake (42.69°N, 120.55°W), in south-central Oregon, sits within the greater ancient Lake Chewaucan basin, a large pluvial lake system, with a proposed highstand at ~13ka (Licciardi, 2001; Negrini, 2002). At its peak, Lake Chewaucan covered 1244 km², was 114 m deep and filled four sub-basins (Licciardi, 2001). Summer Lake sits in the northwest end of the sub-basins and is connected to the Upper Chewaucan Marsh by an overflow channel in its southeastern corner. Further southeast the Chewaucan Basin encompasses Lake Abert and the Lower Chewaucan Marsh as well (see Figure 1 and Appendices, Figure 1A). The Chewaucan Basin has been the subject of numerous studies of past lake history (e.g. Cohen *et al.*, 2000; Negrini *et al.*, 2000; Licciardi, 2001; Zic *et al.*, 2002). Researchers have taken advantage of the extensive tephra layers, well-preserved sediments and fossils. In particular, Licciardi (2001) constrained the timing of the lake-surface fluctuations during the latest Pleistocene lake cycle. I will compare my ages from ²³⁰Th-U geochronology to the ¹⁴C dated lake level record produced by Licciardi (2001).

2.3 ²³⁰Th-U Geochronology

Relative ages of surfaces have successfully been 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 CaCO₃ accumulation (eg. Harden *et al.*, 1991; Tackman, 1993; Redwine, 2003). However, the optimum age dating technique will provide numerical ages. Absolute age-dating techniques include radiocarbon geochronology, ²³⁰Th-U geochronology and the use of cosmogenic isotopes (such as ³⁶Cl, ¹⁰Be, ²⁶Al, ³He) (Benson, 2003). This study used ²³⁰Th-U geochronology of tufas to provide absolute ages of surfaces.

Uranium, a radioactive element, consisting of 99.28 % ²³⁸U ($t_{1/2}$ = 4.46 Gyr), 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 ²³⁴Th and ²³⁴Pa daughter isotopes are short-lived compared to ²³⁸U, ²³⁴U and ²³⁰Th. Samples can 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 ²³⁰Th ($t_{1/2}$ = 75.4 kyr) and ²³⁴U ($t_{1/2}$ = 245 kyr) ²³⁰Th-U geochronology has an

upper age limit of about 500 ka. The equation for calculating the ingrowth of ^{230}Th is (Kaufman and Broecker, 1965):

$$\frac{^{230}\text{Th}}{^{234}\text{U}} = \frac{1 + e^{-\lambda_0 t}}{^{234}\text{U} \backslash ^{238}\text{U}} + \left(1 - \frac{1}{^{234}\text{U} \backslash ^{238}\text{U}}\right) \left(1 - \frac{\lambda_0}{\lambda_0 - \lambda_4}\right) [1 - e^{-(\lambda_0 - \lambda_4)t}] \quad (1)$$

where λ_0 and λ_4 are decay constants for ^{230}Th and ^{234}U , respectively, t is the time since the mineral formed and Th and U are the concentrations of the given Th or U isotope. Under oxidizing conditions typical of surface waters, U is mobile as the hexavalent uranyl ion (U(VI)O_2^{2+}) and is readily incorporated into secondary minerals such as carbonates (Ku, 2000; Reeder *et al.*, 2001). In contrast, Th is relatively insoluble under oxic conditions, and ideal samples have low initial Th concentrations such that all ^{230}Th is derived from radioactive decay of ^{234}U in the sample. However, by measuring the abundance of ^{232}Th in the sample, corrections can be made for any “detrital Th” (e.g. Th derived from other processes aside from purely radioactive decay) that may have originally been incorporated into the mineral (Edwards *et al.*, 1987; Ludwig and Titterton, 1994; Neymark and Paces, 2000; Sharp *et al.*, 2003; Paces *et al.*, 2004; Maher *et al.*, 2007; Fletch *et al.*, 2010).

3. METHODS

3.1 Field Methods

3.1.1 Elevation Transects

Elevation transects were obtained with GPS measurements up gently sloping hills formed primarily on basalt dip slopes and rhyolite plugs (see Figure 2 for location of study area and transects). This location, in the middle of Surprise Valley, has extensive, laterally continuous shoreline features that can be correlated among different hillslopes (Figure 4B). Lake levels were characterized by identifying the edge of horizontal terraces below a crest or rise in topography. Figure 4B and 4C show horizontal terraces, assumed to indicate the erosion of bedrock via wave action at different lake water levels. Eight transects were carried out (for locations see Figure 2B) using a handheld Garmin Oregon 550t GPS unit to measure the distance from origin and elevation of each identified shoreline feature. In addition to the handheld GPS measurements, one very accurate continuous transect was obtained from USGS scientist Jonathan Glen. This transect, originally walked to measure near surface magnetic anomalies simultaneously with GPS co-ordinates, captured the shoreline set from near valley bottom at 1402m to 1470m in the vicinity of transects 3, 4 and 8 (see Figure 2B for location).

Figure 4. Tufa Sample and Field Images. (A) Scanned cut cross section of example tufa sample collected *in situ* on basalt bedrock. This is sample SVT18-1, elevation 1552 m just below the highstand elevation of 1565.5 m. (B) Cropped Digital Orthophoto Quadrangles imagery derived from aerial photos. Laterally continuous shoreline features (terraces and beach ridges) are visible among multiple dip slopes. The location of transects 1 and 5 are denoted by the dashed line. (C) Side view of transect no. 5 standing on shoreline no. 3. Lake levels were characterized by identifying the edge of horizontal wave-cut terraces below a rise in topography associated with beach ridges.

3.1.2 Sample Collection

Tufa samples were collected *in situ* from exposed bedrock on the front edge or crest of the wave-cut terraces and are assumed to correspond to lake levels at that terrace elevation. Eighteen tufa samples were collected in Surprise Valley in the same area as the elevations transects (Figure 2B). All samples were laminar and encrusted on exposed bedrock (Figure 4A). Tufa is not present at all shoreline elevations, and varies in thickness (0.1-5cm) and abundance. Six of eleven shorelines included tufa clasts on the slopes along transects 1, 2 and 5, and all of these were sampled. ^{230}Th -U geochronology is optimal for lake tufas, such as those found on the exposed bedrock below the wavecut surfaces at Surprise Valley, CA. ^{230}Th -U geochronology can also be applied to soil carbonates from the soils developed on surfaces (e.g. Sharp *et al.*, 2003), such as the horizontal shoreline surfaces of shoreline sets found at Summer Lake, OR (Figure 4C).

Hydrothermal carbonate deposits can be found around Surprise Valley corresponding to modern and extinct hot springs (Egger and Miller, 2011), but these were avoided. Sampling of lacustrine tufa was not carried out in the vicinity of these hydrothermal deposits to avoid potential sampling tufa that may have been diagenetically altered.

Additionally, three tufa samples (from elevations of 1333, 1326, and 1319 meters) were collected from the Chewaucan Basin, at the north end of modern Summer Lake (See Figure A1 in Appendices for sampling locations).

3.2 Laboratory Methods

The following is a summary of the laboratory work carried out on the tufa samples. This included the application of ^{230}Th -U geochronology (column chemistry and isotopic analysis), stable isotopic analyses and elemental analyses. In addition, a thorough calibration of a new $^{229}\text{Th}/^{236}\text{U}$ spike (StUTh1), used for ^{230}Th -U geochronology, was carried out before the analysis of samples.

3.2.1 ^{230}Th -U Geochronology

The primary purpose of this study was to date the tufa collected from both basins using the absolute dating technique of ^{230}Th -U geochronology. I processed 21 samples and two aliquots of secular equilibrium opal standard (BZVV from Virgin Valley, NV obtained from the Royal Ontario Museum) using the following methods to accurately determine the Th and U isotopic ratios of the samples. Following this, the two detrital correction calculation methods (TSD and BSE) were applied to the measured isotope ratios of each set of samples.

3.2.1.1 Spike Calibration

In order to accurately measure the U and Th concentrations of carbonates low in U and Th, such as the tufas from Surprise Valley and the Chewaucan Basin, a new isotopic spike optimized for measurement by multi-collector inductively coupled plasma mass spectrometer (MC-ICP-MS) was calibrated. Mixed spikes used in ^{230}Th -U geochronology use isotopes of Th and U (^{229}Th and ^{236}U) not found naturally. ^{236}U and ^{229}Th are both long-lived byproduct isotopes collected during nuclear reactor energy production and from spent nuclear fuel.

Several parameters had to be evaluated in order to dilute the existing spike to the appropriate concentrations of ^{229}Th and ^{236}U for carbonate rocks. The analytical and laboratory constraints that had to be considered were: (1) the average concentration of U and Th in both young and secular equilibrium carbonate samples likely to be analyzed at the Stanford ICP-MS/TIMS facility; (2) the sensitivity and limits of accuracy on measurements made using the Faraday detectors and ion counters on the Nu Plasma High Resolution MC-ICP-MS; (3) the optimal spike to sample ratio of $^{229}\text{Th}/^{230}\text{Th}$; and (4) the amount of spike per sample that can be most precisely measured gravimetrically (found

to be 10-100 μ L). All of these constraints were accounted for when making StUTh1, the mixed spike used in this study.

Following dilution of the spike, the concentrations of ^{236}U and ^{229}Th were determined by using certified gravimetric standards for ^{238}U and ^{232}Th (U standard was CRM 145 from the New Brunswick Laboratory of the US Department of Energy and Th standard was Ames Th Metal from the Ames Laboratory of the US Department of Energy). Neither gravimetric standard has ^{236}U or ^{229}Th .

The final ^{236}U concentration, analyzed over the period of sample analysis, based on an error weighted average of the data (see Appendices, Table A2, Table A3, Figure A2 and Figure A3) was found to be 0.91987 ± 63 ng/g (2σ , MSWD = 0.60, $n = 18$), and the ^{229}Th concentration was found to be 0.22418 ± 58 ng/g (2σ , MSWD = 0.89, $n = 32$). All error propagation used two sigma percent errors and the errors on the concentration measurements were calculated as the square root of the sum of the analytical and gravimetric errors squared.

3.2.1.2 Sample Dissolution

Tufa samples were cut using a micro-saw into 0.5-3 gram chips. Thick laminar samples were targeted to avoid areas of high porosity or possible re-crystallization. To remove dust and clean the tufa, 1-3 gram of sample was rinsed in ethanol and dried.

The clean samples, selected under an optical microscope, were combined with the calibrated mixed spike of ^{236}U and ^{229}Th , and then dissolved in 14 N nitric acid (HNO_3) and treated with hydrogen peroxide (H_2O_2) to remove detrital organic material. Some samples did not fully dissolve and contained an insoluble fraction. The insoluble fraction was dissolved using aqua regia ($\text{HNO}_3\text{-HCl}$), followed by a hydrofluoric acid (HF) digestion. In preparation for chemical purification samples were brought up in 4 mL of 1.5 N HNO_3 . Two complete procedural blanks were prepared in tandem with sample dissolution.

3.2.1.3 Column Chemistry

The dissolved tufa samples were purified by ion exchange chromatography using Eichrom TRU-spec resin in 0.2 mL Teflon columns following the procedure of Luo *et al.*, (1997). The column chemistry recipe is outlined in the Appendices, Table A2. An additional step of 0.6M HF was included at the end of U collection to ensure maximum U

collection. For samples with known concentration and the two aliquots of BZVV, the yield of the column chemistry was found to be ~85 % for U and ~75 % for Th. Along with each set of samples, four in total, a column blank was put through the same chemistry as all samples.

3.2.1.4 Isotopic Analyses

The U and Th column fractions were evaporated to dryness and re-dissolved in 2 % HNO₃ and 2 % HNO₃ + trace HF, respectively. I analyzed the U and Th isotopic compositions within the ICP-MS/TIMS Facility at Stanford University, using a Nu Instruments Plasma HR multi-collector inductively coupled plasma mass spectrometer (MC-ICP-MS) equipped with 12 Faradays and 4 discrete dynode secondary electron multipliers (SEM). 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. All U and Th isotopic standards and solutions were also diluted in 2 % HNO₃ and 2 % HNO₃ + trace 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. U isotopes were measured in static mode with ²³⁶U and ²³⁴U measured simultaneously in the ion counters IC0 and IC1, and ²³⁵U and ²³⁸U were measured on Faraday cups. The measured ²³⁵U/²³⁸U 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 ²³⁴U/²³⁸U and ²³⁶U/²³⁸U ratios for Faraday/IC gain. A certified isotopic standard, CRM 145 (also known as CRM-112) was analyzed repeatedly during the analytical sessions to assess reproducibility. During the time period of the analyses and spike calibration the measured ²³⁴U/²³⁸U was 0.000052613 ± 41 (2 σ , n=37) for CRM 145. The certified ²³⁴U/²³⁸U of CRM 145 is 0.000052841 ± 79 (2 σ).

Th isotopes were measured dynamically with ²²⁹Th and ²³⁰Th in IC0 and ²³²Th in Faraday cups. Bracketing standards of IRMM-035 were used to correct samples for instrumental mass fractionation and Faraday/IC gain. The tail contribution from ²³²Th onto ²³⁰Th was monitored by concurrently measuring the signal at 0.5 and 0.4 amu on the high mass side of the ²³⁰Th peak center and 0.5 amu on the low mass side of the ²³⁰Th peak. By statically measuring 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 ²³²Th

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 correction was between 0.5 % and 1.5 % of the ^{230}Th signal. The $^{229}\text{Th}/^{230}\text{Th}$ ratio was not corrected for mass fractionation, which is expected to be small over this mass range. Th Ames Metal was analyzed repeatedly during the analytical sessions to assess 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 0.00005744 ± 44 (n =34).

The isotopic measurements of secular equilibrium opal standard, BZVV, gave average activity ratios of $(^{234}\text{U}/^{238}\text{U}) = 1.000 \pm 0.003$ and $(^{230}\text{Th}/^{238}\text{U}) = 0.984 \pm 0.035$, for the two samples, as expected. This indicated that the methods used for analytical corrections, column chemistry and the calibrated spike concentrations give true isotopic values for samples processed by the methods outlined above. Procedural blanks, (including dissolution) and column blanks measured using the Nu Attom ICP-MS indicated that procedural blanks were 0.68 pg of U and 3.81 pg of Th, and column blank blanks contributed 1.73 pg of U and 3.67 pg of Th.

All error propagation used two sigma percent errors and error on the individual ratios was calculated as the sum of the internal and external errors as presented above. Final activity ratios used the decay constants of Cheng *et al.* (2000). Age calculations were done using the program Isoplot (Ludwig, 2003a,b). Typically samples with $(^{230}\text{Th}/^{232}\text{Th}) < 20$ require a detrital Th correction to be applied to the age. The Total Sample Dissolution (TSD hereafter) approach (Bischoff and Fitzpatrick, 1991; Luo and Ku, 1991; Ku *et al.*, 1998; Ku, 2000) has previously been applied to samples with $(^{230}\text{Th}/^{232}\text{Th}) < 20$. The U and Th isotopic compositions of at least three coeval samples are plotted on two isochrons (e.g. $(^{234}\text{U}/^{232}\text{Th})$ vs. $(^{230}\text{Th}/^{232}\text{Th})$ and $(^{238}\text{U}/^{232}\text{Th})$ vs. $(^{230}\text{Th}/^{232}\text{Th})$) to obtain the detrital-corrected $(^{230}\text{Th}/^{238}\text{U})$ and $(^{234}\text{U}/^{238}\text{U})$ value from the slope of the linear least squares regression. Geochronologic studies of tufa (Lin *et al.*, 1996; Garnet *et al.*, 2003) and evaporites (Ku *et al.*, 1998) employed this approach. The second widely used approach is to assume that the sample's initial $(^{230}\text{Th}/^{232}\text{Th})$ $(^{234}\text{U}/^{232}\text{Th})$ and $(^{238}\text{Th}/^{232}\text{Th})$ are equal to those of either (1) a second, measured and related sample, such as pure detritus, or (2) an assumed isotopic composition such as the average crustal $^{232}\text{Th}/^{238}\text{U}$ (Ludwig and Titterton, 1994). For this study the latter was

used, and a Bulk Silicate Earth (BSE hereafter) detrital correction was applied by assuming a $^{230}\text{Th}/^{232}\text{Th}$ ratio equal to that of bulk silicate earth ($4.46 \times 10^6 \pm 2.23$ (2σ)) and a $^{232}\text{Th}/^{238}\text{U}$ of 3.8 ± 1.9 (2σ). See Appendices, Section 7.3 for additional explanation of the BSE and TSD detrital corrections, as well as the derivation of equations for the TSD method.

3.2.2 $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and Sr/Ca Analyses

Both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ were measured against a known carbonate standard, Vienna PeeDee Belemnite (hereafter denoted as VPDB). Measurements were made at the Stable Isotope Biogeochemistry Laboratory at Stanford University. Using 0.05-0.1 mg of crushed sample, the 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 internal precision of the instrument is <0.06 ‰ for $\delta^{18}\text{O}$ and <0.03 ‰ for $\delta^{13}\text{C}$ analyses.

Sr and Ca concentrations in carbonate were measured on a Thermo Scientific ICAP 6300 Duo View Spectrometer equipped with a Solid State CID Detector at the Environmental Measurements 1 Laboratory at Stanford University. Following the procedures outlined by Davis *et al.*, (2009), 10-100 mg of crushed sample were digested in 14 N 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). 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 ($R^2 = 0.99$). Finally, the analyses of error associated with sample preparation blanks (Ca, $2\sigma = 0.17$ %; Sr, $2\sigma = 0.40$ %), and analytical error derived from known 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.

4. RESULTS

4.1 Elevation Transects

Where tufa samples were collected, elevation data constrain the lake levels. In addition, multiple transects were carried out to more accurately measure the laterally continuous elevations of lake levels in Surprise Valley. This method was of limited utility because the handheld GPS does not accurately measure elevation. However, the handheld GPS does record latitude-longitude accurately (within 5 to 8 m). Thus, instead of using the raw GPS elevations, the latitude-longitude waypoints were imported into ArcGIS and I extracted elevations using a digital elevation model (DEM) by pinning the latitude-longitude co-ordinates to the DEM surface. The elevations extracted from the continuous transect, T-USGS, obtained from Dr. Jonathan Glen correlated well with the average elevation measurements from the handheld GPS/DEM method. Table 1 and Figure 5 summarize the eleven major shoreline elevations identified as laterally continuous among the eight GPS transects (T1 to T8), and the continuous transect (T-USGS) conducted as part of this study (location of transects are given in Figure 2B).

Despite the limitations of the handheld GPS used in this study, the elevations obtained by pinning the latitude-longitude coordinates to a DEM appear to have reproduced elevations found by previous work and for the continuously measured transect. A previous study by Zimbelman *et al.* (2009), using the shoreline features at the south end of Surprise Valley as an analog for potential pluvial landforms observed on Mars as a means calibrate proposed satellite laser altimeters, identified major shoreline features at 1428 m, 1448 m, 1458 m and 1476 m, and measured the highstand at 1533.6 m. Elder and de la Fuente (2009) estimated a similar highstand elevation of 1530 m at the north end of Surprise Valley by analyzing modern day Lake Annie. This highstand elevation is lower than that found in this study (1565.5 m) and reported by Reheis (1999b) at 1567 m. While this apparent difference in elevation could be due to isostatic rebound of the crust, such as that observed (22 m) to south at the much larger ancient Lake Lahontan (Adams *et al.*, 1999), the elevations recorded by this study and Zimbelman *et al.*, (2009) appear to correlate well. Zimbelman *et al.*, (2009) have identified elevations near or within error of shoreline numbers 2, 4, 5, 6 and 9 measured for this study (see Table 1).

Isostatic rebound would increase elevation measurements made in the center of the valley (this study) more than the measurements made at the north (Elder and de la

Fuente, 2009) and south (Zimbelman *et al.*, 2009) of Surprise Valley. Preliminary modeling of the isostatic rebound of the lake indicate that it would require a very thin, elastic lithospheric plate (~5 km) to produce the >30 m of isostatic rebound required to reproduce the discrepancy (Egger and Huerta, personal communication, 2011), which is unlikely. Additionally, late Pleistocene to Holocene fault motion may have offset the shoreline sets from different basalt dip slopes (Oldow and Singleton, 2008). For the purposes of this study, elevation measurements made in the center of Surprise Valley in the accommodation zone, will be used to assess lake level fluctuations in tandem with ²³⁰Th-U ages presented in the following sections.

Table 1. Average Identified Shoreline Elevations. Using the elevation transects (found in Appendices Tables A1-8 and plotted in Figure 2) the following eleven shorelines were identified. Shorelines 10 and 11 have no average or standard deviation because they were only observed at the top of transect 1.

Shoreline	Elevation Range (m)	Average Elevation (m)	Standard Deviation, σ (m)	Elevation from T-USGS (m)
1	1415.0 - 1422.5	1418.2	2.9	1418.069
2	1425.4 - 1433.3	1429.3	3.0	1423.661
3	1438.6 - 1443.3	1440.6	1.7	1439.522
4	1446.7 - 1451.5	1448.7	2.0	1448.822
5	1457.4 - 1461.0	1458.9	1.6	1454.636
6	1468.4 - 1474.4	1471.5	2.2	1470.353
7	1491.1 - 1493.8	1492.4	1.9	--
8	1513.2 - 1515.3	1514.1	1.0	--
9	1528.2 - 1532.4	1530.3	3.0	--
10	1552.2	--	--	--
11	1565.6	--	--	--

Figure 5. Surprise Valley Elevation Transect Data. Raw elevation and distance from origin from the eight elevation transects used in this study to identify laterally continuous paleo lake levels. Our highest elevation, shoreline number 11 at 1565.6m, agrees with the elevation reported by Reheis (1999b). Standard deviations (denoted by range of grey bars) are 1σ , and questions marks indicate possible additional shoreline elevations.

4.2 ^{230}Th -U Geochronology

The measured U and Th isotopic ratios of all samples are presented in Table 2. The ^{238}U and ^{232}Th concentrations of the tufa samples ranged from 0.15 to 5.50 ppm and 0.188 to 1.373 ppm respectively, with ($^{230}\text{Th}/^{232}\text{Th}$) of 1.7 to 4.5. To accurately obtain ages on samples with a ($^{230}\text{Th}/^{232}\text{Th}$) < 20, the Total Sample Dissolution (TSD) and Bulk Solid Earth (BSE) detrital Th corrections were applied to all measurements. Figure 6 shows the graphical representation of the TSD detrital correction calculation using two plots (e.g. ($^{234}\text{U}/^{232}\text{Th}$) vs. ($^{230}\text{Th}/^{232}\text{Th}$) and ($^{238}\text{U}/^{232}\text{Th}$) vs. ($^{230}\text{Th}/^{232}\text{Th}$)) for each set of apparently coeval samples from Surprise Valley (as outlined in Section 3). Table 3 summarizes the detrital corrected isotope ratios calculated from the slopes of the graphs plotted in Figure 6, and the TSD detrital corrected ages. For comparison, Figure 7 shows a ^{230}Th -U isochron diagram with isotope evolution curves and ages for both detrital correction methods.

From Surprise Valley six samples (denoted SVT#, where # is the sample number), with three coeval subsamples as replicates, were each analyzed to produce ^{230}Th -U age using both detrital correction methods. The TSD detrital correction calculation was not successful with SVT2 because the ($^{230}\text{Th}/^{238}\text{U}$) > 1, which precludes calculation of an age (($^{230}\text{Th}/^{238}\text{U}$) $\approx$ 0.1-0.4 for young samples such as this). This may be due to preferential U loss during open system recrystallization. Otherwise, the subsampling of SVT2 may not have been of coeval samples. SVT2 produced BSE detrital corrected ages of 23.61 ± 4.90 ka and 36.69 ± 4.88 ka. For the ages calculated using the BSE detrital correction an error weighted mean age was calculated (Table 2). Comparing the ages obtained from the two methods of dating indicate that the BSE detrital correction (Table 2) and the TSD detrital correction (Table 3) were within error for five of the six samples analyzed SVT4 (31.13 ± 3.98 ka vs. 30.80 ± 3.53 ka), SVT3 (31.17 ± 3.06 ka vs. 32.29 ± 12.48 ka), SVT14 (22.38 ± 4.6 ka vs. 23.07 ± 2.99 ka) and SV17 (4.64 ± 1.22 ka vs. 1.75 ± 2.43 ka). For the highest elevation sample, SVT18 the ages produced by the two detrital correction methods disagree considerably (13.93 ± 1.21 ka for the BSE detrital correction and 21.99 ± 5.45 ka for the TSD detrital correction).

While the ages between the two detrital correction calculations methods are within error of each other, the range in calculated ($^{234}\text{U}/^{238}\text{U}$)_{initial} values (and the measured ($^{234}\text{U}/^{238}\text{U}$)) varies considerably using both methods (1.217 ± 0.168 to 1.964 ± 0.134 for the TSD method and 1.253 ± 0.014 to 1.945 ± 0.101 for the Bulk Silicate Earth

method). Additionally, using the BSE detrital correction method appears to substantially reduce the two-sigma error.

Figure 7 demonstrates that the ($^{234}\text{U}/^{238}\text{U}$) measurements are segregated into two groups, indicating that unless the ($^{234}\text{U}/^{238}\text{U}$) of the lake water dramatically changed between these time intervals, recrystallization likely occurred when water levels were above previously deposited tufa deposits resulting in open system behavior.

From the Chewaucan Basin three samples from three shoreline elevations (1333 m, 1326 m, 1319 m) were analyzed to produce ^{230}Th -U ages using the BSE detrital correction method (see Table 2). The three detrital corrected ages are within error of each other and increase in age with increased elevation (24.66 ± 13.61 ka, 19.28 ± 8.53 ka and 19.07 ± 6.54 ka) recording the recession of lake level. Additionally, they have similar ($^{234}\text{U}/^{238}\text{U}$)_{initial} ratios (1.897 ± 0.245 , 2.149 ± 0.219 and 2.154 ± 0.167) indicating closed system evolution, in contrast to several of the Surprise Valley samples.

Table 2. ^{230}Th -U Geochronology Measurements and Bulk Silicate Earth (BSE) Detrital Correction Ages. Detrital corrections were made assuming $^{230}\text{Th}/^{232}\text{Th}$ ratio equal to that of bulk silicate Earth ($4.46 \times 10^{-6} \pm 2.23$ (2σ)) and a $^{232}\text{Th}/^{238}\text{U}$ of 3.8 ± 1.9 (2σ). The weighted mean was taken for each bulk sample's coeval replicates. All age calculations and weighted means were made using decay constants from Cheng *et al.*, (2000) and calculated using Isoplot (Ludwig, 2003a;b). Tufa samples from Surprise Valley (SV) and the Chewaucan Basin's Summer Lake (SL) are denoted using the following convention: SVT#-subsample. The number (#) is the sample number and the subsamples are the coeval samples taken to apply the TSD detrital correction calculation. (Table continued on next page with BSE detrital corrected ages and error weighted mean ages)

Location & Sample name	sample weight (g)	Spike Weight (g)	^{232}Th (ppm)	^{238}U (ppm)	Measured Activity Ratios ^{a,b,c}			Uncorrected Age ^{d,e}	
					$(^{230}\text{Th}/^{232}\text{Th})$	$(^{230}\text{Th}/^{238}\text{U})$	$(^{234}\text{U}/^{238}\text{U})$	Uncorrected Age (ka)	$(^{234}\text{U}/^{238}\text{U})_{\text{initial}}$
BZVV 1	0.5031	0.0162			3812.23 ± 76.33	0.979 ± 0.033	1.000 ± 0.003		
BZVV 2	0.5035	0.0150			3997.46 ± 83.54	0.990 ± 0.036	1.000 ± 0.003		
<i>Surprise Valley Samples</i>									
SVT4-1a	1.6057	0.0170	0.855	1.27	2.360 ± 0.019	0.508 ± 0.006	1.481 ± 0.014	44.56 ± 0.85	1.546 ± 0.016
SVT4-1b	1.6057	0.0450	0.539	1.03	2.911 ± 0.026	0.487 ± 0.007	1.513 ± 0.007	41.22 ± 0.74	1.576 ± 0.008
SVT4-1c	1.6057	0.0316	1.373	1.35	1.698 ± 0.027	0.553 ± 0.025	1.370 ± 0.013	54.72 ± 3.19	1.433 ± 0.015
SVT3-1a ^e	1.2881	0.0235	0.417	0.77	2.928 ± 0.179	0.508 ± 0.047	1.511 ± 0.008	43.46 ± 4.83	1.578 ± 0.012
SVT3-1a-Thdup	1.2881	0.0235	0.400	0.77	2.988 ± 0.027	0.497 ± 0.007	1.511 ± 0.008	42.32 ± 0.76	1.576 ± 0.009
SVT3-2a	2.0287	0.0383	0.688	1.22	2.844 ± 0.028	0.514 ± 0.008	1.537 ± 0.015	43.13 ± 0.92	1.607 ± 0.016
SVT3-3a	1.1865	0.0221	0.446	0.67	2.410 ± 0.085	0.513 ± 0.028	1.466 ± 0.008	45.64 ± 3.01	1.531 ± 0.010
SVT2-2a ^f	1.0069	0.0134	0.359	0.70	2.562 ± 0.024	0.421 ± 0.006	1.575 ± 0.007	33.20 ± 0.57	1.632 ± 0.008
SVT2-1a ^f	1.2243	0.0152	0.318	0.64	3.496 ± 0.028	0.556 ± 0.007	1.582 ± 0.008	45.78 ± 0.74	1.663 ± 0.008
SVT14-1a-1	0.2919	0.0247	0.225	0.24	1.707 ± 0.018	0.516 ± 0.008	1.656 ± 0.005	39.51 ± 0.75	1.734 ± 0.006
SVT14-1a-2	0.5016	0.0136	0.188	0.29	2.241 ± 0.021	0.471 ± 0.007	1.732 ± 0.005	33.75 ± 0.60	1.805 ± 0.006
SVT14-1c ^e	0.6356	0.0189	0.756	0.40	1.902 ± 0.017	1.157 ± 0.016	1.732 ± 0.005	108.34 ± 2.38	1.994 ± 0.009
SVT17-1a-1 ^e	0.8913	0.0689	0.277	1.46	1.755 ± 0.018	0.106 ± 0.002	1.240 ± 0.004	9.72 ± 0.16	1.246 ± 0.004
SVT17-1a-1-Thdup	0.8913	0.0689	0.281	1.46	1.735 ± 0.018	0.107 ± 0.002	1.240 ± 0.004	9.76 ± 0.17	1.246 ± 0.004
SVT17-1a-2	1.3491	0.0490	0.592	4.20	1.815 ± 0.015	0.082 ± 0.001	1.244 ± 0.004	7.36 ± 0.12	1.249 ± 0.004
SVT17-1b ^e	0.1067	0.0186	0.095	0.15	1.486 ± 0.015	0.292 ± 0.004	1.283 ± 0.004	27.84 ± 0.47	1.306 ± 0.004
SVT18-1d ^g	1.8064	0.0465	0.599	5.26	3.187 ± 0.106	0.116 ± 0.006	1.254 ± 0.012	10.50 ± 0.58	1.262 ± 0.012
SVT18-1c	1.7331	0.0457	0.666	5.09	4.533 ± 0.050	0.189 ± 0.003	1.235 ± 0.012	18.03 ± 0.39	1.247 ± 0.013
SVT18-1e	2.6690	0.0704	0.947	5.50	3.350 ± 0.065	0.184 ± 0.006	1.242 ± 0.012	17.35 ± 0.68	1.255 ± 0.012
SVT18-1e-Thdup	2.6690	0.0704	0.878	5.50	3.522 ± 0.043	0.179 ± 0.003	1.242 ± 0.012	16.88 ± 0.39	1.254 ± 0.012
<i>Chewaucan Basin Samples</i>									
SLT1-2c-1 ^f	0.9022	0.0112	0.503	0.38	1.417 ± 0.012	0.597 ± 0.009	1.543 ± 0.007	51.51 ± 0.94	1.628 ± 0.008
SLT2-1 ^f	1.3738	0.0178	0.426	0.41	1.582 ± 0.017	0.526 ± 0.009	1.787 ± 0.008	36.89 ± 0.71	1.874 ± 0.009
SLT3-1b ^f	1.0957	0.0181	0.269	0.32	1.812 ± 0.018	0.490 ± 0.007	1.847 ± 0.008	32.71 ± 0.58	1.929 ± 0.009

Table 2. (Continued)

Location & Sample name	Detrital Corrected Age			Weighted Mean Age
	Corrected Age (ka)	(²³⁴ U/ ²³⁸ U) _{initial}	rho	Corrected Age (ka)
BZVV 1				
BZVV 2				
<i>Surprise Valley Samples</i>				
SVT4-1a	31.05 ± 6.90	1.640 ± 0.074	-0.07	31.14 ± 3.98
SVT4-1b	31.10 ± 5.29	1.651 ± 0.055	-0.05	
SVT4-1c	31.63 ± 12.50	1.556 ± 0.111	-0.03	
SVT3-1a ^c	32.95 ± 7.33	1.656 ± 0.059	0.05	31.17 ± 3.06
SVT3-1a-Thdup	28.85 ± 5.10	1.819 ± 0.148	-0.46	
SVT3-2a	32.37 ± 5.57	1.693 ± 0.065	-0.10	
SVT3-3a	32.16 ± 7.53	1.621 ± 0.070	-0.03	
SVT2-2a ^f	23.61 ± 4.90	1.713 ± 0.059	-0.02	30.18 ± 83.14
SVT2-1a ^f	36.69 ± 4.88	1.745 ± 0.058	-0.13	
SVT14-1a-1	22.26 ± 8.49	1.934 ± 0.158	-0.10	22.48 ± 4.60
SVT14-1a-2	22.56 ± 5.47	1.945 ± 0.101	-0.10	
SVT14-1c ^c	75.53 ± 37.04	2.839 ± 0.790	-0.82	
SVT17-1a-1 ^c	5.22 ± 2.38	1.256 ± 0.010	0.16	4.64 ± 1.22
SVT17-1a-1-Thdup	5.19 ± 2.42	1.256 ± 0.010	0.16	
SVT17-1a-2	4.05 ± 1.75	1.256 ± 0.008	0.15	
SVT17-1b ^g	13.18 ± 7.93	1.351 ± 0.041	0.14	
SVT18-1d ^{c,g}	7.86 ± 1.57	1.268 ± 0.014	0.03	13.93 ± 1.21
SVT18-1c	14.94 ± 1.87	1.253 ± 0.014	0.00	
SVT18-1e	13.30 ± 2.36	1.264 ± 0.015	0.04	
SVT18-1e-Thdup	13.13 ± 2.14	1.263 ± 0.015	0.03	
<i>Chewaucan Basin Samples</i>				
SLT1-2c-1 ^f	24.66 ± 13.61	1.897 ± 0.243	-0.12	
SLT2-1 ^f	19.28 ± 8.53	2.149 ± 0.219	-0.13	
SLT3-1b ^f	19.07 ± 6.54	2.154 ± 0.167	-0.12	

^a Brackets denote activity ratio

^b Decay constants used are from Cheng et al. (2000)

^c Error is 2s (absolute)

^d Uncorrected age, not corrected for detrital ²³⁰Th contribution

^e Excluded from isochrons for corrected ages based on inconsistent uncorrected age or re-measurement of Th isotopes

^f Isochron approach not feasible

^g Excluded from weighted mean age

Figure 6. Isochron Plots Calculating Detrital Corrected Activity Ratios. The TSD detrital correction calculation was applied to five samples from Surprise Valley (SVT4, SVT3, SVT14, SVT17 and SVT18). The values for the slopes (equations for U and W) of $(^{230}\text{Th}/^{234}\text{U})_{\text{authigenic}}$ and $(^{234}\text{U}/^{238}\text{U})_{\text{authigenic}}$ are calculated using a least squares regression. Similar to Lou and Ku (1991) and Ku *et al.* (1998), the least squares fit does not consider error-weighting and correlation of the measured activity ratios. The least square regressions and error calculations were carried out in Isoplot (Ludwig, 2003a;b).

Table 3. Total Sample Dissolution (TSD) Detrital Correction Activity Ratios and Ages. TSD calculated detrital corrected activity ratios $(^{230}\text{Th}/^{234}\text{U})_{\text{authigenic}}$ and $(^{234}\text{U}/^{238}\text{U})_{\text{authigenic}}$ are calculated from the slope of a linear least squares regression (Figure 6) for each set of coeval replicates. $(^{230}\text{Th}/^{238}\text{U})_{\text{authigenic}}$ is calculated by multiplying the two detrital corrected activity ratios together. These activity ratios are then used to calculate a detrital corrected age. The weighted linear least squares regression and age calculates were carried out on Isoplot (Ludwig, 2003a;b).

Isochron Calculated Activity Ratios ^{a,b}						Detrital Corrected Age ^b		
			^(230Th/234U) authigenic	^(234U/238U) authigenic	^(230Th/238U) authigenic	Age (ka)	^(234U/238U) initial	rho
Shoreline No.	Elevation (m)							
SVT4	2	1429	0.194 ± 0.013	1.903 ± 0.130	0.369 ± 0.035	30.80 ± 3.53	1.713 ± 0.105	-0.57
SVT3	3	1441	0.261 ± 0.051	1.560 ± 0.300	0.407 ± 0.112	32.29 ± 12.48	1.614 ± 0.316	-0.54
SVT14	6	1472	0.251 ± 0.013	1.653 ± 0.100	0.415 ± 0.033	23.07 ± 2.99	1.964 ± 0.134	-0.55
SVT17	8	1514	0.016 ± 0.022	1.227 ± 0.024	0.020 ± 0.027	1.75 ± 2.43	1.228 ± 0.024	0.05
SVT18	10	1552	0.184 ± 0.022	1.204 ± 0.160	0.222 ± 0.040	21.99 ± 5.45	1.217 ± 0.168	-0.59

^a Brackets denote activity ratio

^b $(^{230}\text{Th}/^{234}\text{U})_{\text{authigenic}}$ and $(^{234}\text{U}/^{238}\text{U})_{\text{authigenic}}$ are calculated as the slope of measured $(^{238}\text{U}/^{232}\text{Th})$ vs $(^{234}\text{U}/^{232}\text{Th})$ and $(^{234}\text{U}/^{232}\text{Th})$ vs. $(^{230}\text{Th}/^{232}\text{Th})$ respectively. These plots are shown in Figure 6.

^c Decay constants used are from Cheng et al. (2000)

Figure 7. Isochrons ($(^{230}\text{Th}/^{238}\text{U})$ vs. $(^{234}\text{U}/^{238}\text{U})$) for Surprise Valley Tufa Samples. (A) BSE detrital corrected $(^{230}\text{Th}/^{238}\text{U})$ and $(^{234}\text{U}/^{238}\text{U})$ for each subsample. The ages of these subsamples produced one age for the BSE detrital correction using an error weighted mean calculation using Isoplot (Ludwig, 2003a;b). (B) TSD detrital corrected $(^{230}\text{Th}/^{238}\text{U})$ and $(^{234}\text{U}/^{238}\text{U})$ for each set of coeval samples used to calculate a detrital corrected age (Table 3). Figures were made using Isoplot (Ludwig, 2003a;b). The data points are color coded by sample (SVT4 = yellow, SVT3 = green, SVT2 = orange, SVT14 = red, SVT17 = blue, SVT18 = black). The TSD detrital correction was not applied to SVT2 due to $(^{230}\text{Th}/^{238}\text{U}) > 1$, calculated from the slopes (as calculated for other samples in Figure 6).

4.3 $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and Sr/Ca Analyses

Eleven samples from Surprise Valley were analyzed for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ and the results of these analyses are presented in Table 3. The $\delta^{13}\text{C}$ varied from -6.142‰ to 4.103‰ and $\delta^{18}\text{O}$ varied from -12.559‰ to -2.501‰ (VPDB).

Table 4. Summary of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ Analyses for Surprise Valley Tufa Samples.

Sample ID	Shoreline No	Mass Analyzed (mg)	$\delta^{13}\text{C}_{\text{VPDB}} (\text{‰})$	$\delta^{18}\text{O}_{\text{VPDB}} (\text{‰})$
SVT4-1a	2	0.099	3.866	-3.123
SVT4-1b	2	0.096	3.472	-3.385
SVT3-1a	3	0.095	3.282	-4.045
SVT3-2	3	0.105	3.302	-2.988
SVT2-Inn	4	0.089	4.103	-2.501
SVT2-Out	4	0.068	3.598	-2.851
SVT14-1a	6	0.073	3.789	-3.238
SVT14-1c	6	0.070	3.690	-3.225
SVT17-1a	8	0.067	-6.218	-10.814
SVT17-1b	8	0.053	-4.748	-11.493
SVT18-1e	10	0.083	-6.142	-12.559

In general $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ are more depleted with increasing elevation (Figure 8A). Additionally, the relationship of $\delta^{18}\text{O}$ vs. $\delta^{13}\text{C}$ for each data point (Figure 8B) demonstrates that they covary with a high correlation coefficient ($R^2 = 0.979$).

Eleven samples were measured for Ca and Sr concentrations in order to obtain Sr/Ca ratios. Sr concentrations vary from 1.01 to 59.60 $\mu\text{g/mL}$, Ca concentrations vary from 14.56 to 223.73 mg/mL and 1000*Sr/Ca vary from 0.64 to 3.56 ($\mu\text{g/mg}$).

Table 5. Summary of Sr and Ca Analyses for Surprise Valley Tufa Samples. Total analytical error on the Sr/Ca ratios is 2.73 % (2σ).

Sample ID	Shoreline No	Mass Analyzed		1000*Sr/Ca	
		(g)	Sr ($\mu\text{g/mL}$)	Ca (mg/mL)	($\mu\text{g/mg}$)
SVT4-1a	2	0.0220	39.35	152.43	2.58
SVT4-1b	2	0.0125	22.55	87.77	2.57
SVT3-1a	3	0.0207	18.12	68.91	2.63
SVT3-2	3	0.0295	59.60	223.73	2.66
SVT2-Inn	4	0.0090	16.88	69.97	2.41
SVT2-Out	4	0.1036	17.26	65.53	2.63
SVT14-1a	6	0.0072	15.51	43.54	3.56
SVT14-1c	6	0.0057	11.07	31.09	3.56
SVT17-1a	8	0.0062	2.05	32.27	0.64
SVT17-1b	8	0.0032	1.01	14.56	0.70
SVT18-1e	11	0.0370	26.14	168.03	1.56

Sr/Ca ratios decrease with increasing shoreline elevation (Figure 8C). Additionally, the relationship of $\delta^{18}\text{O}$ vs. Sr/Ca for each data point (Figure 8D) demonstrates that the two geochemical proxies correlate well ($R^2 = 0.720$) with $\delta^{18}\text{O}$ being more depleted and Sr/Ca ratios decreasing with increasing shoreline elevation.

Figure 8. $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and Sr/Ca Measurements from Surprise Valley Tufa. Stable and elemental measurements obtained on tufa samples from Surprise Valley. (A) $\delta^{18}\text{O}$ (blue circle) and $\delta^{13}\text{C}$ (red square) measurements vs. tufa sample elevation. (B) $\delta^{18}\text{O}$ (blue circle) and Sr/Ca (green triangle) measurements vs. tufa sample elevation. (C) $\delta^{18}\text{O}$ vs. $\delta^{13}\text{C}$ plot showing strong correlation ($R^2 = 0.979$) between the measurements. (D) $\delta^{18}\text{O}$ vs. Sr/Ca plot showing correlation ($R^2 = 0.720$) between the measurements. Linear least square regressions used to fit data to test for correlation.

5. DISCUSSION

5.1 Assessing Surprise Valley as a Terminal Basin

The extent to which the Surprise Valley lake system was a terminal basin, with no outflow or substantial regional groundwater flow in the past, is important for using lake shoreline ages (and other geochemical data) as indicators of past climate. Smaller, closed basin lakes are more likely to respond rapidly to watershed scale changes in precipitation, runoff/infiltration and evaporation (Menking *et al.*, 2004; Garcia and Stokes, 2006). Coupling light stable isotopic and elemental analysis with geochronologic data provides tools to assess whether Lake Surprise was a closed, inward draining pluvial lake as suggested by modern watershed data (Figure 2B). The covariance of $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ -Sr/Ca can also be used to evaluate the role of evaporation in lake systems (Muller *et al.*, 1972; Eugster and Kelts, 1983; Talbot and Kelts, 1990; Davis *et al.*, 2009).

Terminal basin lakes have long 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 and Kelts, 1990). Evaporative effects on $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ indicate that if a lake is hydrologically closed, and evaporation is relatively high (as observed in Figure 3B for Surprise Valley's modern climate), the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ will covary (Davis *et al.*, 2009; Talbot and Kelts, 1990). Correlation of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ demonstrated in Figure 8 shows that with increased shoreline elevation both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ are more depleted, consistent with a closed basin (Figure 8A). Additionally, as lake level rose and fell the values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ remained well correlated, $R^2 = 0.979$ (Figure 8C).

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. Additionally, in terminal lake systems, Sr is not removed as efficiently from the lakewater as Ca, which is taken up by precipitating carbonates, thus Sr is concentrated when evaporation, 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 out of the system. At Surprise Valley, increased Sr/Ca ratios at lower lake levels (Figure 8B) are well correlated with the increased relative evaporative effects recorded in the $\delta^{18}\text{O}$, $R^2 = 0.720$ (Figure 8D), again consistent with a closed basin.

However, the isotope and elemental data may be biased towards high correlation coefficients by the clustering of the data (Figure 8C and D). The lower elevation cluster (top right corner) give correlation coefficients of $R^2 = 0.43$ for $\delta^{18}\text{O}$ vs. $\delta^{13}\text{C}$, $R^2 = 0.30$ for $\delta^{18}\text{O}$ vs. Sr/Ca. A larger set of samples (ideally from more shoreline elevations) is necessary to further quantify the evaporative effects at Lake Surprise. The expansion of this dataset, if a hydrologically open system is suspected, will elucidate the contributing effects of local hot springs and/or fluxes between the lake and groundwater aquifers (Li *et al.*, 2008a,b).

In addition to the geochemical data, the stream data derived from high-resolution elevation data obtained from NASA's Shuttle Radar Topography Mission, EPA watershed demarcations, and the contoured outline of the Reheis (1999b) highstand, all shown in Figure 2A, demonstrate that topographically the Surprise Valley lake system likely remained a terminal basin throughout its late Pleistocene pluvial history. Additionally, while some overflow into Duck Flat appears to have occurred at the highstand elevation, to the south, the smaller region of overflow also appears to remain a closed based on the outline of the Reheis (1999b) highstand elevation and local topography (Figure 2B). Coupling of the geochemical measurements and geophysical observations supports previous hypotheses (Russell, 1927; Sack, 2002) suggesting that the Surprise Valley lake system remained a terminal basin in the late Pleistocene.

5.2 Lake Levels of Late Pleistocene to Middle Holocene Lake Surprise

Linking ^{230}Th -U geochronology of tufas with elevation data allows for the hydrologic changes of a lake system to be assessed directly by constructing a lake level curve (Figure 9). Rather than using indirect proxies for lake level, such as pollen (e.g. Whitlock *et al.*, 1999; Grigg *et al.*, 2001; Grigg and Whitlock, 2002) or stable isotopes in sedimentary carbonates from lake cores, the absolute dating of materials collected at known elevations provide temporal constraints on the fluctuation of lake systems through time. Corrections for isostatic rebound may be necessary (as discussed previously). However, for smaller lake systems such as Lake Surprise, the correction is likely minimal (Egger and Huerta, personal communication, 2011).

Figure 9 illustrates the lake levels of Lake Surprise from the late Pleistocene to middle Holocene. Both methods of detrital Th correction are plotted in Figure 9, the BSE age calculations yield lower age errors, are internally consistent, and appear to accurately

correct for relative amounts of detrital Th incorporated into the sample. In the lower elevation samples, both detrital Th correction methods produced ages within error for the same suite of sample measurements. At the highest elevation sample and SVT17, there is a discrepancy of ages between the two detrital Th correction methods. Using the TSD method, the lake level curve suggests an unlikely scenario of lake level doubling (1472 m to 1552 m) within only ~ 1 kyr. In contrast, using the BSE detrital Th corrected ages, the doubling of lake level would have occurred over 9.5 kyr during the period from 22.48 ± 4.60 ka to 13.93 ± 1.21 ka, over the middle to latter part of the LGM. The closest age to modern obtained in this study (using the BSE detrital correction method) was 4.64 ± 1.22 ka at 142 m above the modern valley bottom. Thus, while the transgression of lake level up 1552 m was recorded temporally in this study, the regression of the lake surface down to modern day levels remains unrecorded beyond the knowledge that the lake was relatively high until 4.64 ± 1.22 ka.

In addition, the data suggests that Lake Surprise was present at moderate water levels (69 to 57 m deep) well before the LGM, from at least 31.17 ± 3.06 ka. Hence, Lake Surprise was likely present well into the later stages of the last interglacial period (Marine Isotope Stage 3), which was apparently much colder and wetter compared to the modern day interglacial.

The three tufa samples dated from the middle shoreline elevations of the Chewaucan Basin (Figure 10 (h)) all yielded ages within error of each other (24.66 ± 13.61 ka, 19.28 ± 8.53 ka, and 19.07 ± 6.54 ka). However, the ages do not agree with the published lake level curve produced by Licciardi (2001). Using sediment cores and dating of shoreline materials Licciardi (2001) produced a preliminary curve with high uncertainty during the LGM, the age of the three samples dated in this study. The Chewaucan Basin's lake history is well constrained by ^{14}C dating approaches from the latest highstand at ~ 13 ka to the early Holocene, but earlier lake levels during the LGM are poorly constrained (Licciardi, 2001; Negrini, 2002). However, these ages place new constraints on the previously hypothesized (Licciardi, 2001; Negrini, 2002) pluvial period occurring throughout much of MIS 2, and prior to the 13 ka highstand.

Figure 9. Lake Surprise Lake Level Curve. ^{230}Th -U ages calculated using both detrital Th correction methods (Tables 2 and 3, and Figure 7) versus the average elevation of the shoreline that the tufa sample was collected (Table 1 and Figure 5). The highstand elevation observed in this study and reported by Reheis (1999b) is given as the dashed line while the lowest modern valley elevation, found in the middle playa lake, is represented by the dash and dotted line.

5.3 Comparison of Results to Existing Lake Records and Other Paleoclimate Proxy Records

The high lake level recorded in Surprise Valley at 13.93 ka occurs within error of the Chewaucan Basin stillstand (13.04 ka) to the north (Licciardi, 2001; Negrini, 2002) and after the Lake Lahontan highstand (15.22 ka) to the south (Adams and Wesnousky, 1998; Godsey *et al.*, 2011). The data presented in this study suggests that Lake Surprise remained at a moderate lake level during the LGM, with the lake level peaking after the LGM, similar to the Chewaucan Basin. Table 6 summarizes lake highstands, significant stillstands and hydrologic maxima from western North America. Hydrologic maxima were defined as periods of time when known high lake level occurred in lakes occupying open basins with potential spillover (e.g. Lake Bonneville), or where lake level broadly peaks over longer time periods (e.g. Babicora Basin, Searless Lake and Lake Manly). Most lake highstands in the Basin and Range region occur in the latter portion of the LGM or after the LGM, during the deglaciation period. Lake highstand ages listed in Table 6 (in approximate north to south order) show an approximate northward migration in the timing of the highstands during the LGM deglaciation.

Table 6. Western North America Lake Stand Ages. Summary of chronologic data compiled from the shoreline studies. All radiocarbon dates (^{14}C Before 1950) have been converted to calendar ages (ka) in order to compare within high stand ages, lake level curves and ^{230}Th -U ages from studies using ^{230}Th -U geochronology.

Lake, State	Late Pleistocene Lake Stand Ages (ka $\pm$ 1 σ)	Sources
Chewaucan Basin, OR	Higher Highstand? $\approx$ 18.1 to 15.7 Highstand - $13.04 \pm .69$ Lower Stillstands - 24.66 ± 13.61 , 19.28 ± 8.53 and 19.07 ± 6.54	Licciardi (2001); Negrini (2002) <i>(This study)</i>
Surprise Valley, CA	Stillstands - 31.17 ± 3.98 to 22.48 ± 4.60 Highstand - 13.93 ± 1.21 Stillstand - 4.64 ± 1.22	<i>This study</i>
Lake Bonneville, UT	Hydrologic maximum from 17.57 to 14.45	Oviatt <i>et al.</i> , (1992); Godsey <i>et al.</i> , (2011)
Lake Franklin, NV	Highstand - 19.09 ± 0.15 Stillstand 1 - 15.52 ± 0.17 Stillstand 2 - 14.82 ± 0.15	Lillquist (1994)
Lake Lahontan, NV	Highstand - 15.223 ± 1.20	Adams and Wesnousky, 1998)
Diamond Valley, NV	Highstand - 26.54 ± 0.27 Stillstand - 15.72 ± 0.23	Tackman (1993)
Newark Valley, NV	Highstand - 16.024 ± 1.15	Redwine (2003); Kurth <i>et al.</i> , (2011)
Jakes Lake, NV	Highstand - 16.16 ± 0.13 Stillstand 1 - 15.73 ± 0.12 Stillstand 2 - 14.35 ± 0.13	Garcia and Stokes (2006)
Lake Manly, CA	Hydrologic maximum from 24.7 to 16	Ku <i>et al.</i> , (1998)
Searless Lake, CA	Hydrologic maximum from 22.29 to 16.5	Smith (1984)
Lake Estancia, NM	Highstand 1 - 27.66 ± 0.14 Highstand 2 - 21.16 ± 0.13 Stillstand 1 - 16.14 ± 0.14 Stillstand 2 - 14.94 ± 0.13	Allen and Anderson (2000); Anderson <i>et al.</i> (2002)
San Agustin Basin, NM	Highstand - 22.6 to 25	Markgraf <i>et al.</i> (1984)
Lower Playas Valley, NM	Highstand - 18.2	Ballenger <i>et al.</i> (2011)
Cloverdale, NM	Highstand - 21.5 to 23.9	Krider (1998)
Babicora Basin, Mexico	Hydrologic maximum from 38 to $\sim$ 29ka	Metcalf <i>et al.</i> (2002)

In addition to the apparent northward migration of highstands during the deglaciation, many of the lake record curves (Figure 10) indicate that lake levels were higher before or during the LGM relative to the Holocene or interglacial (MIS 3), followed by a pulse or peak in lake surface height in the latter stages of the LGM or after the LGM. The well-dated records from Lake Bonneville, Lake Lahontan, Searles Lake and Lake Surprise demonstrate this trend. Many authors have postulated a southward deflection of the jet stream and westerly storm tracks with increasing growth of the Laurentide and Cordilleran ice sheets (Negrini, 2002; Benson *et al.*, 1990; Adams and Wesnousky, 1998; Zic *et al.*, 2002). Interpretations of several speleothem records, such as

the Cave of the Bells record (Wagner *et al.*, 2010) presented in Figure 10 have also invoked the shifting westerly winds as the atmospheric connection between increased rainfall and increased ice sheet extent (Oster *et al.*, 2009; Asmeron *et al.*, 2010).

In addition, southward shifted westerlies in the Northern Hemisphere have been proposed to mirror changes in the Southern Hemisphere westerlies and the inter-tropical convergence zone via synchronous shifts in the three primary circulation cells of Earth's idealized atmospheric circulation model: the Hadley, Ferrel, and Polar cells (Toggweiler *et al.*, 2006). Air masses flowing from ca. 30°N/S (high pressure) to ca. 60°N/S (low pressure) in the Ferrel cell are deflected by the Coriolis effect to produce the prevailing westerly winds observed in the mid-latitudes of the Earth's atmosphere (Toggweiler, 2009). Mid-latitude moisture availability is dominated by the westerly wind patterns during Northern Hemisphere winters and is the primary moisture source for pluvial lakes in the Basin and Range. Modern remote sensing analysis and compilation of paleoclimate data indicate that westerly wind patterns and their associated storm tracks correlate with latitudes of maximum precipitation (Moreno *et al.* 1999). Satellite observations and models of future climate (e.g. Shindell and Schmidt, 2003) have also shown a poleward shift of the westerlies in both hemispheres, due to global temperature increases (Salathe, 2006). Furthermore, Southern Hemisphere paleoceanographic records and ice cores suggest southward shifted, strengthened Southern Hemisphere westerlies beginning at 17 ka (Toggweiler *et al.*, 2006; Lamy *et al.*, 2007). Additionally, marine radiocarbon evidence suggests that it was this shift in the strength and position of the westerlies in the Southern Hemisphere that enhanced ventilation of the deep ocean and thus led to increasing atmospheric CO₂ (Marchitto *et al.*, 2007), creating a positive feedback loop to warm the Earth and bring it out of the LGM (Toggweiler and Russel, 2008).

Linking south shifted westerlies during glacial periods with a symmetric northward response during deglaciation is attractive because of its synchronous, apparently insolation forced explanation. Denniston *et al.*, (2007) even call for the synchronicity of climate changes between the western United States and the high-latitudes to explain the transmission of far-field atmosphere-ocean interactions in the poles and tropics to temperatures in the Basin and Range. However, with respect to explaining the lag of lake highstands behind peaks in soil P-ET (Maher *et al.*, in prep) (Figure 10) this hypothesis may be too simplistic, neglecting the important effect that decreased temperature and decreased relative humidity may have on evaporation.

Decreased evaporation would have the same apparent effect of increases in rainfall brought by south shifted westerly storm tracks; particularly in evaporation dominated arid environments, such as Surprise Valley (Figure 3A), surface evaporation should be diminished due to decreased temperatures and increased relative humidity.

Ultimately, ice sheet extent is driven by changes in solar insolation (Figures 10f and 10g) resulting in changes in atmospheric CO₂ levels via atmosphere-ocean-biosphere feedback cycles (Denton *et al.*, 2010). Moreover, analysis of globally widespread paleoclimate datasets indicates that the temperature conditions at the Earth's surface are also driven by solar insolation, particularly during glacial cycles (Denton *et al.*, 2010). Although this forcing may have lag times, the changes in temperature recorded in the NGRIP $\delta^{18}\text{O}$ temperature proxy (Rasmussen *et al.*, 2006), Devil's Hole $\delta^{18}\text{O}$ temperature proxy (Winograd *et al.*, 2006) and alkenone Sea Surface Temperature (SST) records (Liu *et al.*, 2005) indicate that the Northern Hemisphere temperatures during the periods before, during and after the LGM follow winter insolation (Figure 10 (a), (c), (d) and (f)). Additionally, new soil records calculating changes in precipitation minus evapotranspiration (P-ET), or soil infiltration rates (Maher *et al.*, in prep) indicate that moisture availability appeared to synchronously peak in the Great Basin and Mojave deserts before the Last Glacial Maximum (Figure 10 (e)), preceding most of the lake record highstands (Table 6) by 8-10 kyr. This suggests that if the Northern Hemisphere westerlies mirrored the Southern Hemisphere westerlies, shifting poleward from 19 to 17 ka (mirroring ice sheet volume), then North America lake highstand peaks, occurring during and after this time period, cannot be solely explained by the passing westerlies hypothesis (Benson *et al.*, 1990; Negrini, 2002). Thus, colder temperatures and decreased summer solar insolation following the LGM may have resulted in the lake highstands, rather than large increases in precipitation.

Figure 10. Compilation of Regional Western North America Paleoclimate Records. Comparison of multiple paleoclimate records and proxies with published lake level curves from the Basin and Range and the Lake Surprise lake level curve produced as part of this study. Lake level curves (right panel) are arranged approximately north to south. (a) NGRIP Greenland ice core $\delta^{18}\text{O}$ on extended GICC05 timescale (Rasmussen *et al.*, 2006), red line is smoothed at 1 %; (b) $\delta^{18}\text{O}$ from Cave of the Bells (Wagner *et al.*, 2010); (c) vein calcite $\delta^{18}\text{O}$ from Devils Hole, Nevada (Winograd *et al.*, 2006); (d) Alkenone SST proxy from Ocean Drilling Program sites 1012 (Liu *et al.*, 2005) and 2020 (Liu, 2004); (e) reconstructed relative $\Delta\text{P-ET}$ (%) from Maher *et al.*, (in prep) pink, Newark Valley; dark blue, Diamond Valley; light blue, Fish Lake Valley; purple, Yucca Mountain. (f) January and July insolation for 40°N calculated using the program of (Laskar *et al.*, 2004); (g) SPECMAP $\delta^{18}\text{O}$ record (Martinson *et al.*, 1987); (h) Chewaucan Basin relative lake level curve from sediment cores (Licciardi, 2001) and $^{230}\text{Th-U}$ dated tufa samples (this study); (i) Lake Surprise lake level fluctuations from $^{230}\text{Th-U}$ dated tufa samples (this study), BSE detrital corrected ages only; (j) and (k) Lake Lahontan, NV and Lake Bonneville, UT relative lake levels (Compiled by Brook *et al.*, 2006, from McCoy, 1987; Oviatt *et al.*, 1987; Benson *et al.*, 1990; Oviatt *et al.*, 1992); (e) Lake Manly, CA lake level (m) (Ku *et al.*, 1998); (f) Searles Lake, CA relative lake level (Smith, 1984); (i) Babicora Basin, Mexico relative lake level (Metcalf *et al.*, 2002). The radiocarbon dated lake level curves (h), (j), (k), (i), (m) and (n)) have substantial error and uncertainty due to the necessary ^{14}C dead carbon correction and the radiocarbon to calendar age conversion uncertainty, particularly after the LGM due to plateaus in the calibration curve. The average standard error in the lake level curves is 1 to 3 ka (error bars given on figure).

The nature of paleoclimatic records make it fundamentally difficult to separate the effects of decreased temperatures versus increased precipitation using lake shoreline records (Reheis, 1999a). Modeling of the Estancia Basin lake system hydrology (highstands and stillstands listed in Table 6) concluded that rainfall might have doubled relative to modern values to produce large pluvial lakes (Menking *et al.*, 2004). Conversely, pollen records from California suggest changes in precipitation of less than 50 % (Lyle *et al.*, 2010).

The best apparent explanation for post-LGM lake highstands such as Surprise Valley is that during a period of decreased summer insolation, lake levels increased due to decreased summer temperatures and reduced surface evaporation. Lake levels are primarily controlled by runoff within a lake's watershed and surface evaporation (Reheis, 1999a). While lakes are likely recording summer conditions (that of decreased temperature), speleothems, groundwater systems and soils appear to reflect winter temperature conditions. As a result, the apparent northward migration of pluvial maxima during the last deglaciation track the receding Laurentide and Cordilleran ice sheets, with highstands also occurring during this period due to a minima in summer insolation.

Thus, the direct link between south-shifted westerlies and ice sheet extent, while not necessarily at odds with this data synthesis, was not the primary control on lake levels. Instead, south-shifted westerlies may have been simply a contributing factor to increasing moisture availability during this period of pluvial events. Further work to quantify and model the watershed scale hydrologic processes (e.g. Menking *et al.*, 2004), regional to global scale atmospheric connections that lead to increased moisture availability (e.g. Kim *et al.*, 2006), and lake evaporation fluxes (e.g. Gibson and Edwards, 2002) would help to clarify the fundamental question of whether decreased temperature or increased precipitation contributed to the expansive pluvial systems observed in the lake records during the period of the last deglaciation.

5.4 Global Climate Models and Paleoclimate Records

In addition to providing a predictive tool for future climate scenarios due to anthropogenic CO₂ emissions, climate models can also be used to evaluate hypotheses for Earth's past climate. This includes an assessment of the variability of the mid-latitude jet stream over North America that may have brought moisture to the now arid Basin and Range. Early modeling results focused on the late Pleistocene suggest that during the LGM both precipitation and westerly winds were enhanced over North America due to a stronger Aleutian Low over the Northeast Pacific Ocean (Kutzbach *et al.*, 1988). Additionally, more recent modeling has shown that ice sheet extent, both continental ice sheets and sea ice, influence the strength and position of westerly storm tracks (Bromwich *et al.*, 2004; Sewall and Sloan, 2004). Conversely, paleoclimate proxies, such as P-ET from uranium isotopes in soil minerals (Maher *et al.*, in prep) or sea surface temperature (SST) from alkenones in ocean sediment cores (as applied in Liu *et al.*, (2005)), can be used to directly test the magnitude and direction of temporal changes predicted by the climate model outputs (Booth *et al.*, 2006; Braconnot *et al.*, 2012).

The Paleoclimate Model Intercomparison Project 2 (PMIP2) (Braconnot *et al.*, 2007) has compiled a database of the results from published model simulations. Figure 11 shows the LGM anomaly (LGM minus modern) for P-ET, precipitation, mean annual wind vectors and zonal wind speed at both 200 millibar and 850 millibar from the Hadley Center Coupled Model, version 3 (HadCM3) run for the LGM and pre-industrial conditions (Gordon *et al.*, 2000). This model is one of many used in the past and current rounds of Intergovernmental Panel on Climate Change (IPCC) assessment reports (Solomon *et al.*, 2007; Braconnot *et al.*, 2007). All parameters were imported as model outputs except for evaporation. The evaporation parameter required calculation, using the relationship between the latent heat flux (a model output) and the mass of water evaporated.

The LGM anomaly maps in Figure 11 demonstrate that HadCM3 does predict southward shifted wind patterns during the LGM, particularly in the upper (the 200 millibar) atmosphere level (Panels B and D). For Surprise Valley, the model predicts approximately double the rainfall amount (Figure 11A, ~500 mm/yr increase with a modern average of 566 mm/yr (Table A1)). However, P-ET (Figure 11B) predicted by HadCM3 is at least two orders of magnitude greater than those predicted by the uranium isotope soil infiltration proxy (Figure 10 (e)).

Figure 11. LGM Anomaly Maps for HadCM3 - 200 and 850 millibar Winds (m/s), Precipitation (mm/yr) and Precipitation Minus Evapotranspiration (mm/yr). HadCM3 is an AOGCM coupling sea-ice, atmospheric and ocean parameters with an atmospheric grid resolution of $3.75^\circ \times 2.5^\circ \times 19$ vertical layers developed by the Hadley Centre in the United Kingdom. The HadCM3 modelling experiments, from ref. (Gordon *et al.*, 2000) are available on the PMIP2 database (Braconnot *et al.*, 2007; <http://pmip2.lsce.ipsl.fr/pmip2/>). All maps are annually averaged anomaly maps, LGM minus modern.

The second set of LGM anomaly maps shown in Figure 12 illustrate additional climate parameters applicable to the paleoclimate records summarized in Figure 10. Analysis of total cloud cover data from HadCM3 demonstrate that cloud cover over the Basin and Range was $\sim 20\%$ greater during the LGM (Figure 12A), indicating increased humidity and a net increase in albedo. While we cannot directly measure past cloud cover, this increase over the Basin and Range is consistent with moisture requirements needed to sustain pluvial lakes during the LGM.

HadCM3 predicts 0 to 200 mm/yr increase in evapotranspiration in the Basin and Range during the LGM (Figure 12B). In comparison, modern evaporation was calculated as 1178 mm/yr (Table A1) using the Kimberly Penman procedure (Dockter, 2004). Additionally, it is expected that during the LGM evapotranspiration would decrease, not increase as predicted by this model. This discrepancy (both in direction and magnitude of change for evaporation) in effectively modeling evapotranspiration may explain why

HadCM3 over-predicts P-ET predicted by orders of magnitude relative to U isotopes in soils (Figure 12C and D). While HadCM3 does not directly output model evapotranspiration in the PMIP2 database, it does have a soil evaporation term. Combined with precipitation, this term should be similar to the uranium isotope proxy illustrated in Figure 10 (e) (Maher *et al.*, in prep).

Figure 12. LGM Anomaly Maps for HadCM3 - (A) Total Cloud Cover (%), (B) Evaporation (mm/yr), (C) Precipitation Minus Soil Evaporation (mm/yr) and (D) Precipitation Minus Evapotranspiration (mm/yr). For details on the HadCM3 AOGCM model see Figure 11.

The HadCM3 model does predict LGM temperatures accurately (Gordon *et al.*, 2000; Braconnot *et al.*, 2007), yet the parameterization of the hydrologic cycle over western North America (and likely in most of the model) clearly requires major refinement. This observation, a major discrepancy between the paleoclimate record and the paleoclimate model's hydrologic quantities, is similar to observations made by a more thorough assessment of the global water cycle in the PMIP2 models, the Climate Model Intercomparison Project (CMIP) 5 project, used in the IPCC Assessment Report 4 (Solomon *et al.*, 2007) to assess future warming scenarios (Waliser *et al.*, 2007). Analysis by Waliser *et al.*, (2007) of historical data sets for precipitation, cloud water, precipitable

water, runoff, soil moisture and snow mass indicate that while the current generation of models are predicting precipitation reasonably well at a global scale, the other remaining hydrologic quantities, particularly snow mass, runoff and soil moisture, remain highly variable among climate models. Furthermore, on a regional scale there exists a large discrepancy between the models (Waliser *et al.*, 2003). Additional work is clearly necessary to improve AOGCMs' parameterization of the fundamental quantities relevant to the Earth's hydrologic balance. The newest suite of PMIP3/CMIP5 models, currently underway, will require mass balance between continental river runoff and precipitation (Taylor *et al.*, 2012), previously not required in PMIP2/CMIP3 models. Improvements to quantify terrestrial components of the models to accurately track soil moisture, soil and lake evaporation, river runoff and groundwater recharge/discharge, will go a long way in producing AOGCMs with the predictive power to inform policy decisions in the future.

6. CONCLUSION

Evaluating the conditions required to fundamentally change atmosphere-lithosphere-hydrosphere interactions during periods of Earth's past poses a challenge that necessitates the integration of geochemical, geochronologic and geophysical tools with numerical modeling approaches. To evaluate the factors that controlled pluvial lake systems in the Basin and Range during the LGM requires a spatially and temporally diverse set of paleoenvironmental/paleohydrologic data to infer paleoclimatic change. Often paleoclimate records rely on proxies requiring accurate calibration and modern analogs (e.g. Oster *et al.*, in press; Maher *et al.*, in prep). However, the approach taken in this study was more direct. Using materials found at specific shoreline elevations, the history of lake level fluctuation was evaluated at two study sites.

By dating shoreline elevations using tufa deposits I reconstructed past lake levels in Surprise Valley (Figure 9). Lake Surprise was moderately full (57 m deep, elevation of 1429 m to 1441 m) during the last interglacial, beginning at least ~31 ka, and increased in size between 31.17 ± 3.06 ka to 22.48 ± 4.6 ka (57 m to 100 m deep). Between 22.48 ± 4.6 ka and 13.93 ± 1.21 ka, Lake Surprise nearly doubled in lake depth from 100 m to 180 m deep (up to an elevation of 1552 m). The age from this shoreline is likely very close to the highstand age of latest Lake Surprise lake cycle. Samples analyzed from a shoreline at an elevation of 1514 m yielded a middle Holocene age of 4.64 ± 1.22 ka, indicating that high lake levels at water depths of 140-180 m may have persisted from 13.93 ± 1.21 ka, through the early Holocene until 4.64 ± 1.22 ka.

In comparison, tufa samples from three elevations in the Chewaucan Basin yielded ages of 24.66 ± 13.61 ka, 19.28 ± 8.53 ka, and 19.07 ± 6.54 ka, all older than the previous ^{14}C -dated highstand of 13.04 ka. These ages are likely associated with a previously unknown pluvial period occurring throughout much of MIS 2.

The Surprise Valley lake level curve precedes the existing Chewaucan Basin record (Licciardi, 2001; Negrini, 2002) to the north and follows the Lake Lahontan lake level record and highstand (Adams and Wesnousky, 1998) to the south.

The synthesis of existing paleoclimate records from western North America reveals an interesting interplay between key hydrologic quantities. The question of whether increased precipitation or decreased temperature provided the necessary available moisture to grow and sustain pluvial lakes remains an ongoing area of research, which can potentially be answered by the establishment of a more spatially extensive

network of paleoclimate records. The diverse data analyzed here, consisting of lake records, soil infiltration records, speleothem records and temperature proxies (Figure 10), indicate that lake highstand ages, including Lake Surprise, primarily precede the LGM and lag behind peaks in soil infiltration by 8 to 10 kyr. Previously, the lake highstands were attributed to the passing of the westerly winds. However, inspection of insolation curves, temperature proxies and speleothems indicate that decreased summer insolation and temperature likely decreased summer evaporation, aiding the development of pluvial lake highstands.

Finally, hydrologic quantities from outputs of the HadCM3 AOGCM were used to evaluate challenges and uncertainties in the predictive power of AOGCMs. Most of the models used by the IPCC do appear to predict temperatures accurately in future global warming scenarios and in paleoclimate model simulations (Gordon *et al.*, 2000; Braconnot *et al.*, 2007). However, when categorically different forcings from the LGM are applied, the hydrological outputs appear to accurately predict the direction of change (in precipitation or P-ET), but over predict the magnitude of change over western North America.

This study illustrates some of the challenges associated with, and the importance of, producing paleoclimate records. The utility of paleoclimate data to decipher Earth's past climatic conditions relies on accurate geochronology and ^{230}Th -U dating of materials, such as tufas, to yield absolute ages, allowing the correlation of climate proxies (Figure 10). Additionally, studies occurring during time intervals of abrupt climate change, such as the transitions from glacial to interglacial conditions following the LGM, may clarify the response of the climate system to large perturbations (e.g. Denton *et al.*, 2010). Furthermore, the comparison between AOGCM simulations of the LGM and the paleoclimate data demonstrated that hydrologic quantities such as evapotranspiration, soil moisture and runoff will require additional paleoclimate proxy data to reduce the model uncertainty (Seager *et al.*, 2007; Waliser *et al.*, 2007) associated with the expected future hydrological changes in the arid western United States driven by global climate change.

7. APPENDICES

7.1 Map of Chewaucan Basin, OR

Figure A1. Map of Chewaucan Basin, Oregon Sample Collection Locations. Shaded relief map of the Chewaucan basin (1:400:000) with the Environmental Protection Agency (EPA) watersheds outlined in purple. The location of tufa samples used in this study and sediment cores used by Licciardi (2001) are denoted by black diamonds and black circles respectively. Map layers for watershed areas are established by the EPA (http://cfpub.epa.gov/surf/huc.cfm?huc_code=18080001) and streams have been established by the USGS HydroSHEDS program (<http://hydrosheds.cr.usgs.gov/>). A shaded relief and USGS 100k topography map were used as the basemap.

7.2 Surprise Valley Watershed Statistics

Table A1. Surprise Valley Watershed Climatic and Hydrologic Statistics Watershed size was calculated using ArcGIS layers from the EPA watershed layers. Lake depth and maximum lake size area were calculated by Reheis (1999b) and Clement (2005) intersected this data with regional Digital Elevation Models to calculate the maximum lake volume. Meteorological data from the WRCC (Western Regional Climate Center) is a historical dataset spanning 1884-2011 from the Cedarville, Lake City and Fort Bidwell, CA weather stations. The meteorological data from AgriMet is compiled by the Pacific Northwest Cooperative Agricultural Weather Network database spanning 1985-2011.

Measured/Calculated Hydrologic Parameters	Quantity	Source
Watershed size	2,313 km ²	EPA Watershed Demarcation
Maximum lake depth	1567 m – 1372 m = 195 m	Reheis (1999b) and this study
Maximum lake area	1,478 km ²	Reheis (1999b)
Lake Volume	321 km ³	Clement (2005)

Annual Historical Climate Data	Quantity	Source
Rainfall Total	56.58 cm/yr	WRCC
Pan Evaporation	117.88 cm/yr	AgriMet
Evaporation Total	140.12 cm/yr	AgriMet
Mean Temperature	9.22 °C	WRCC
Mean High Temperature	16.85 °C	WRCC
Mean Low Temperature	1.59 °C	WRCC
Mean Relative Humidity	57.58 %	AgriMet

7.3 Spike Calibration

The following tables and graphs are the measurements made over multiple analysis days to calibrate the mixed spike, StUTh1, containing ^{229}Th and ^{236}U . All of the dilutions were carried out gravimetrically, weighing everything six times. CRM 145 β -dilution has a ^{238}U concentration of 15870.4046 ng/g and Th Ames Metal β -dilution has a ^{232}Th concentration of 924.8870 ng/g. Both standards were spiked with StUTh1 and diluted to concentrations suitable for MC-ICP-MS analysis. This calibration measured on the same aliquot of spike used to spike the tufa samples, locking in the $^{230}\text{Th}/^{238}\text{U}$ ratio prior to sample dissolution, elemental separation and analytical isotope measurement.

Table A2. Calibrated ^{236}U Concentration in StUTh1 ‘Primary.’ Errors include gravimetric and analytical errors. Isotopic standard CRM U010 was used to correct the $^{236}\text{U}/^{238}\text{U}$ ratios for Faraday/IC gain. The ^{238}U concentration of the spiked CRM145 was 15.7451 ng/g. The concentration of ^{236}U was calculated using an error weighted average and was found to be 0.91987 ± 0.00063 ng/g (2σ , MSWD = 0.60, $n = 18$). These measurements are plotted graphically in Figure A2.

Date	$^{236}\text{U}/^{238}\text{U}$	^{236}U Concentration (ng/g)	2σ (%)
May 9 th 2012 ¹	5.912E-05 $\pm$ 1.6E-07	0.9177 $\pm$ 0.0024	0.264
May 9 th 2012 ¹	5.904E-05 $\pm$ 1.5E-07	0.9164 $\pm$ 0.0024	0.261
May 9 th 2012 ¹	5.871E-05 $\pm$ 1.5E-07	0.9114 $\pm$ 0.0023	0.253
May 9 th 2012 ¹	5.884E-05 $\pm$ 1.5E-07	0.9133 $\pm$ 0.0023	0.252
May 9 th 2012 ¹	5.885E-05 $\pm$ 1.5E-07	0.9135 $\pm$ 0.0023	0.256
May 9 th 2012 ¹	5.894E-05 $\pm$ 1.5E-07	0.9148 $\pm$ 0.0023	0.250
May 14 th 2012	5.926E-05 $\pm$ 2.0E-07	0.9198 $\pm$ 0.0031	0.337
May 14 th 2012	5.917E-05 $\pm$ 1.9E-07	0.9185 $\pm$ 0.0030	0.330
May 14 th 2012	5.924E-05 $\pm$ 1.5E-07	0.9196 $\pm$ 0.0024	0.257
May 14 th 2012	5.924E-05 $\pm$ 1.5E-07	0.9195 $\pm$ 0.0024	0.257
May 14 th 2012	5.922E-05 $\pm$ 1.5E-07	0.9193 $\pm$ 0.0023	0.249
May 14 th 2012	5.922E-05 $\pm$ 1.5E-07	0.9192 $\pm$ 0.0023	0.255
May 16 th 2012	5.917E-05 $\pm$ 1.6E-07	0.9184 $\pm$ 0.0025	0.273
May 16 th 2012	5.916E-05 $\pm$ 1.8E-07	0.9183 $\pm$ 0.0028	0.304
May 16 th 2012	5.927E-05 $\pm$ 1.8E-07	0.9201 $\pm$ 0.0028	0.305
May 16 th 2012	5.934E-05 $\pm$ 1.8E-07	0.9211 $\pm$ 0.0028	0.304
May 16 th 2012	5.917E-05 $\pm$ 1.9E-07	0.9185 $\pm$ 0.0030	0.323
May 16 th 2012 ²	5.898E-05 $\pm$ 1.8E-07	0.9155 $\pm$ 0.0029	0.313
May 16 th 2012	5.931E-05 $\pm$ 1.8E-07	0.9207 $\pm$ 0.0028	0.306
May 16 th 2012	5.935E-05 $\pm$ 1.8E-07	0.9212 $\pm$ 0.0028	0.306
May 16 th 2012	5.934E-05 $\pm$ 1.8E-07	0.9211 $\pm$ 0.0028	0.304
May 16 th 2012	5.933E-05 $\pm$ 1.8E-07	0.9210 $\pm$ 0.0028	0.300
May 16 th 2012	5.934E-05 $\pm$ 1.8E-07	0.9211 $\pm$ 0.0028	0.305
May 16 th 2012	5.935E-05 $\pm$ 1.8E-07	0.9213 $\pm$ 0.0028	0.305
May 16 th 2012	5.926E-05 $\pm$ 1.8E-07	0.9199 $\pm$ 0.0029	0.312

¹ Spike-standard equilibration appeared to be occurring during this first day of analysis (May 9th 2012) after the solution was made (Figure A2, Panel A), these measurements were discarded from final weighted average (See Figure A2).

² Observed a mass bias correction issue, caused raw $^{236}\text{U}/^{238}\text{U}$ value of the solution and $^{235}\text{U}/^{238}\text{U}$ value of CRM U010 brackets to fall outside the range of standard error for the day’s analysis, this measurement was discarded from final weighted average.

Figure A2. ^{236}U concentration calibration in StUth1 spike. The horizontal black lines are the error weighted average concentration. 32 measurements were made on the spiked CRM 145 solution during three analytical sessions. (a) The first analytical day was discarded, there appeared to be spike-sample equilibrations problems. On that analytical day CRM 145 was spike at the beginning of the day. (b) Summary of just the measurements used to calculated the error weighted mean concentration. Graph and calculations were made using Isoplot (Ludwig, 2003a,b).

Table A3. Calibrated ^{229}Th Concentration in StUTh1. Errors include gravimetric and analytical errors. Isotopic standard IRMM35 was used to correct the $^{229}\text{Th}/^{230}\text{Th}$ ratios for Faraday/IC gain. The ^{232}Th concentration of the spiked Th Ames Metal was 41.4855 ng/g. The concentration of ^{229}Th was calculated using an error weighted average and was found to be 0.22418 ± 0.00058 ng/g (2σ , MSWD = 0.89, $n=32$). These measurements are plotted graphically in figure A3.

Date	$^{229}\text{Th}/^{232}\text{Th}$	^{229}Th Concentration (ng/g)	
			2σ (%)
April 25 th 2012	2.192E-05 $\pm$ 2.6E-07	0.2226 $\pm$ 0.0026	1.174
April 25 th 2012	2.198E-05 $\pm$ 2.0E-07	0.2231 $\pm$ 0.0020	0.903
April 25 th 2012	2.207E-05 $\pm$ 2.4E-07	0.2241 $\pm$ 0.0025	1.095
April 25 th 2012	2.257E-05 $\pm$ 3.3E-07	0.2292 $\pm$ 0.0033	1.440
May 1 st 2012	2.234E-05 $\pm$ 4.1E-07	0.2269 $\pm$ 0.0042	1.837
May 1 st 2012	2.225E-05 $\pm$ 3.9E-07	0.2259 $\pm$ 0.0039	1.734
May 1 st 2012	2.223E-05 $\pm$ 3.4E-07	0.2257 $\pm$ 0.0034	1.508
May 1 st 2012	2.230E-05 $\pm$ 2.5E-07	0.2265 $\pm$ 0.0025	1.102
May 1 st 2012	2.200E-05 $\pm$ 3.4E-07	0.2233 $\pm$ 0.0034	1.539
May 1 st 2012	2.220E-05 $\pm$ 3.1E-07	0.2254 $\pm$ 0.0032	1.413
May 1 st 2012	2.208E-05 $\pm$ 3.4E-07	0.2241 $\pm$ 0.0035	1.558
May 1 st 2012	2.198E-05 $\pm$ 3.8E-07	0.2232 $\pm$ 0.0039	1.735
May 1 st 2012	2.256E-05 $\pm$ 5.7E-07	0.2290 $\pm$ 0.0058	2.529
May 1 st 2012	2.202E-05 $\pm$ 3.0E-07	0.2235 $\pm$ 0.0031	1.373
May 10 th 2012	2.197E-05 $\pm$ 3.2E-07	0.2231 $\pm$ 0.0032	1.444
May 10 th 2012	2.195E-05 $\pm$ 2.6E-07	0.2229 $\pm$ 0.0027	1.197
May 10 th 2012	2.200E-05 $\pm$ 3.3E-07	0.2233 $\pm$ 0.0034	1.503
May 20 th 2012	2.200E-05 $\pm$ 6.6E-07	0.2233 $\pm$ 0.0067	3.005
May 20 th 2012	2.208E-05 $\pm$ 3.8E-07	0.2242 $\pm$ 0.0038	1.710
May 20 th 2012	2.214E-05 $\pm$ 4.0E-07	0.2248 $\pm$ 0.0040	1.790
May 20 th 2012	2.211E-05 $\pm$ 3.9E-07	0.2245 $\pm$ 0.0039	1.744
May 20 th 2012	2.194E-05 $\pm$ 2.8E-07	0.2227 $\pm$ 0.0029	1.287
May 20 th 2012	2.212E-05 $\pm$ 3.8E-07	0.2246 $\pm$ 0.0038	1.699
May 20 th 2012	2.202E-05 $\pm$ 3.1E-07	0.2236 $\pm$ 0.0032	1.418
May 20 th 2012	2.198E-05 $\pm$ 3.0E-07	0.2232 $\pm$ 0.0031	1.382
May 20 th 2012	2.204E-05 $\pm$ 3.3E-07	0.2238 $\pm$ 0.0033	1.482
May 20 th 2012	2.214E-05 $\pm$ 3.8E-07	0.2248 $\pm$ 0.0039	1.717
May 20 th 2012	2.201E-05 $\pm$ 4.2E-07	0.2234 $\pm$ 0.0043	1.903
May 20 th 2012	2.198E-05 $\pm$ 3.5E-07	0.2232 $\pm$ 0.0035	1.587
May 20 th 2012	2.215E-05 $\pm$ 3.3E-07	0.2249 $\pm$ 0.0033	1.477
May 20 th 2012	2.214E-05 $\pm$ 4.1E-07	0.2248 $\pm$ 0.0041	1.838
May 20 th 2012	2.196E-05 $\pm$ 3.0E-07	0.2230 $\pm$ 0.0030	1.361

Figure A3. ^{236}Th concentration calibration in StUTH1. The horizontal black line is the error weighted average. 32 measurements were made on the spiked Th Ames Metal solution during four analytical sessions. Graph and calculations were made using Isoplot (Ludwig, 2003a,b).

7.4 Column Chemistry Modified Recipe

Table A4. Column Chemistry Procedure 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 200-400 mesh	
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

7.3 Explanation of the Detrital Thorium Age Correction Methods

Below is a derivation of the Total Sample Dissolution (TSD) detrital correction method (cf. Ku, 2000). Equations 2 and 4 describe the linear least squares regression (unweighted as suggested by Ku (2000)). The derived equations 3 and 5 below are the slopes, corresponding to the detrital corrected activity ratios. Equations 4 and 7 are the y intercepts, given a constant y intercept for both equations the linearity is expected of the ($^{234}\text{U}/^{232}\text{Th}$) vs. ($^{230}\text{Th}/^{232}\text{Th}$) and ($^{238}\text{U}/^{232}\text{Th}$) vs. ($^{230}\text{Th}/^{232}\text{Th}$) plots.

$$\frac{^{230}\text{Th}}{^{232}\text{Th}} = U \frac{^{234}\text{U}}{^{232}\text{Th}} + E \quad (2)$$

$$U = \left(\frac{^{230}\text{Th}^*}{^{234}\text{U}} \right)_{\text{authigenic}} \quad (3)$$

$$E = \left(\frac{^{230}\text{Th}}{^{232}\text{Th}} \right)_0 e^{-\lambda_0 t} + \left(\frac{^{234}\text{U}}{^{232}\text{Th}} \right)_{\text{detrital}} \left[\left(\frac{^{230}\text{Th}^*}{^{234}\text{U}} \right)_{\text{detrital}} - \left(\frac{^{230}\text{Th}^*}{^{234}\text{U}} \right)_{\text{authigenic}} \right] \quad (4)$$

$$\frac{^{234}\text{U}}{^{232}\text{Th}} = W \frac{^{238}\text{U}}{^{232}\text{Th}} + F \quad (5)$$

$$W = \left(\frac{^{234}\text{U}}{^{238}\text{U}} \right)_{\text{authigenic}} \quad (6)$$

$$F = \left(\frac{^{238}\text{U}}{^{232}\text{Th}} \right)_{\text{detrital}} \left[\left(\frac{^{234}\text{U}}{^{238}\text{U}} \right)_{\text{detrital}} - \left(\frac{^{234}\text{U}}{^{238}\text{U}} \right)_{\text{authigenic}} \right] \quad (7)$$

The variables U and W (equations 3 and 6) are the slopes of equations 2 and 5, corresponding to the detrital corrected activity ratios ($^{230}\text{Th}/^{234}\text{U}$) and ($^{234}\text{U}/^{238}\text{U}$) used in the age equation (equation 1). E and F (equations 4 and 7) are the y-intercepts of equations 2 and 5, corresponding to the detrital ($^{230}\text{Th}/^{232}\text{Th}$) and ($^{234}\text{U}/^{232}\text{Th}$) incorporated into the mineral at the time of mineral precipitation.

The second approach, Bulk Silicate Earth (BSE), detrital correction used by others is to assume that the sample's initial ($^{230}\text{Th}/^{232}\text{Th}$) ($^{234}\text{U}/^{232}\text{Th}$) and ($^{238}\text{Th}/^{232}\text{Th}$) are equal to those of either (1) a second, measured and related sample, such as pure detritus, or (2) an assumed isotopic composition such as the average crustal $^{232}\text{Th}/^{238}\text{U}$ (Ludwig and Titterton, 1994). For this study, the latter method was applied and a correction was applied assuming a $^{230}\text{Th}/^{232}\text{Th}$ ratio equal to that of bulk silicate earth ($4.46 \times 10^6 \pm 2.23$ (2 σ)) and a $^{232}\text{Th}/^{238}\text{U}$ of 3.8 ± 1.9 (2 σ). The geochronology program Isoplot, developed at the Berkeley Geochronology Center (Ludwig, 2003a,b), has built in functions to calculate corrected ages based on this method.

7.5 Elevation Transect Data Tables

The following tables summarize the data collected by walking transects with handheld GPS. Eight transects (T-1 to 8) were carried out in total. See location of transects on map of study area (Figure 2B). “?” denotes duplicate surfaces observed at similar elevations.

Table A5. Transect 1

Distance (m)	Elevation (m)	Shoreline No.
0.0	1415.0	1
105.7	1428.6	2
177.9	1439.1	3
266.2	1451.5	4
365.3	1457.9	5
438.1	1470.0	6?
528.1	1474.4	6?
661.9	1491.1	7
838.1	1513.2	8?
899.8	1513.8	8?
1011.4	1532.4	9
1152.5	1552.2	10
1207.4	1565.6	11

Table A6. Transect 2

Distance (m)	Elevation (m)	Shoreline No.
0.0	1422.5	1
33.9	1425.4	2
110.8	1440.3	3
221.5	1457.4	4
299.1	1470.7	6

Table A7. Transect 3

Distance (m)	Elevation (m)	Shoreline No.
0.0	1425.4	2?
79.4	1432.8	2?
238.9	1438.6	3
346.2	1447.0	4
430.0	1458.4	5

Table A8. Transect 4

Distance (m)	Elevation (m)	Shoreline No.
0.0	1440.5	3
90.1	1447.7	4
190.5	1448.7	4
376.1	1457.8	5

Table A9. Transect 5

Distance (m)	Elevation (m)	Shoreline No.
0.0	1417.4	1
105.1	1428.6	2?
135.7	1433.3	2?
200.8	1446.7	4
284.1	1461.0	5
340.7	1472.5	6

Table A10. Transect 6

Distance (m)	Elevation (m)	Shoreline No.
0.0	1493.8	7
418.3	1515.3	8
491.6	1520.8	?
592.9	1528.2	9

Table A11. Transect 7

Distance (m)	Elevation (m)	Shoreline No.
0.0	1408.6	Below 1
234.2	1419.7	1
457.4	1431.3	2
580.2	1441.5	3
696.8	1450.6	4
800.7	1460.8	5
869.6	1468.4	6?
896.5	1472.9	6?

Table A12. Transect 8

Distance (m)	Elevation (m)	Shoreline No.
0.0	1402.9	Below 1
265.3	1410.9	Below 1
467.8	1416.6	1
701.6	1428.8	2
829.2	1434.9	?
963.5	1443.3	3

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