Late Glacial and Deglacial Fluctuations of Mono Lake, California

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ABSTRACT

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Anthropogenic climate change risks significant changes in the global distribution of precipitation. Across the western United States, modelling studies show significant reductions in wetness that imply weighty societal and ecological impacts. But the validity of the model projections need to be ground-truthed. Paleo-hydroclimate data are useful reference points to assess a model’s ability to hindcast past hydroclimate. If the hindcast matches the paleodata, it brings confidence to a model’s ability to predict future hydroclimatic change.

The foremost metric of hydroclimate in the geologic record is the surface area of lakes in hydrologically closed basins. In such basins, a lake’s surface area is determined by the balance between precipitation and evaporation. The lake will expand when the balance is positive, and it will contract when the balance is negative.

In this dissertation, I develop a 25-9 ka record of lake fluctuation from the Mono Basin, a hydrologically closed basin in east-central California. I deduced the fluctuations using three pieces of evidence: stratigraphy; geomorphology; and geochronology. These pieces of evidence were determined from a study of the Mono Basin’s Late Pleistocene lithostratigraphic unit: the Wilson Creek Formation.
There are 19 tephra intercalated in the Wilson Creek tephra. They are named by their reverse depositional order (Ash 19 is the oldest and Ash 1 is the youngest). Uncertainty on their ages cause confusion as to the paleo-hydroclimate record of the Mono Basin. The age of Ash 19, for example, is important because its deposition marks the onset of relatively high lake levels that occurred during the last glaciation. There are two principal interpretations of Ash 19’s age: 40 ka, which is based on lacustrine macrofossil $^{14}$C data; and 66 ka, which is underpinned by paleomagnetic intensity data. In chapter 2, I tested these end-member interpretations. I used the U/Th method to date carbonate deposits that underlie and cut across Ash 19. The U/Th data show that Ash 19 must have been deposited between these two dates: $66.8 \pm 2.8$ ka; and $65.4 \pm 0.3$ ka. These dates are, therefore, more consistent with the 66 ka interpretation of Ash 19’s age. Thus the onset of relatively high lake levels in the Mono Basin corresponds with the rapid drawdown of atmospheric CO$_2$ during Marine Isotope Stage 4. The coincidence between the drop in atmospheric CO$_2$ and lake level rise is suggestive of a causal link.

In chapter 3, I determined Mono Lake's fluctuations 25-9 ka. This time encompasses three climatic intervals: the coolest time of the last glaciation, termed the Last Glacial Maximum (LGM); the period corresponding to the rapid termination of the last glaciation, termed the deglaciation; and the early Holocene, a period of inordinate warmth that immediately followed the last glaciation’s termination. In this study, I used stratigraphic and geomorphic evidence in conjunction with $^{14}$C and U/Th dates. I measured the $^{14}$C dates on bird bones and charcoal. And I measured the U/Th dates on carbonates. Together the data showed that the lake's rises and falls concurred with North Atlantic climate. Periods of aberrant warmth in the North Atlantic concurred with low stands of Mono Lake. On the other hand, extreme cooling in the North Atlantic correlated with Mono Lake high stands. The timing of these lake fluctuations also
corresponds with variations in other tropical and mid-latitude hydroclimatic records. The global harmony in the hydroclimatic records suggests a unifying conductor. I hypothesize that the conductor is tropical atmospheric circulation.

In chapter 4, I present evidence on the peculiar case of an extreme low stand of Mono Lake. The low stand is dubbed the “Big Low”. The principal evidence underpinning the Big Low derives from a sedimentary sequence exposed along the canyon walls of Mill Creek. The strata show that the lake fell below 1,982 m between the deposition of Ashes 5 and 4—making this low stand the lowest recognized level of Mono Lake during the Wilson Creek Formation. Observations from dispersed sequences corroborate this interpretation. And three data constrain the age of the Big Low to be between ~24.4-20.5 ka: a carbonate U/Th date on a littoral conglomerate associated with the Big Low; a carbonate U/Th date that underlies Ash 4; and a carbonate U/Th date that cuts across Ash 5. Thus the interval that the Big Low must occur within encompasses the LGM. The timing of this low stand, therefore, corresponds with summer temperature minima, suggesting that the fall was due not to an increase in evaporation but due to a decrease in precipitation. This finding is counter to conventional wisdom: that the LGM was a relatively wet interval. In addition, both the documentation of a low stand during glacial maximum conditions and the inference that precipitation must have been reduced are contrary to previous published interpretations from model and paleoclimatic data. These discrepancies raise significant questions about our understanding of the regional expression and forcing of hydroclimate across the western United States during the LGM. Because of this period’s importance to ground-truthing climatic models, additional evidence on the geographic extent of this unexpected result is essential.
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Dedication

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Chapter 1: Introduction

1.1 Abstract

Anthropogenic climate change risks monumental changes in the global distribution of precipitation. Some regions will get wetter. Others will get drier. Model simulations project significant reductions in western US wetness. These reductions call for the assessment of water resource management strategies. The validity of these projections, however, needs to be evaluated. Paleowetness data are useful references points to assess a model’s ability to retrodict past climate. If the retrodiction matches the paleodata, it brings confidence to a model's ability to predict future climate.

Lake fluctuation records are the preeminent indicators of past hydroclimate. When their basins are hydrographically closed, the balance between precipitation(p) and evaporation(e) determine their fluctuations. The lake rises when p-e increases; and it falls in the converse. Thus a geologic archive of lake fluctuations is an accurate measure of past hydroclimate.

In this doctoral thesis, I show evidence that constrains the fluctuations of Mono Lake 25-9 ka. This interval of time corresponds with the glaciation’s maximum and its rapid collapse as well as the first few kyr of the present interglaciation. The principal underpinning of this work is a stratigraphic investigation of the Mono Basin’s Late Pleistocene lithostratigraphic unit, the Wilson Creek Formation(WCF).

The WCF contains 19 tephra. They are numbered by their reverse depositional order. Ash 1 is the youngest. Ash 19 is the oldest. The age of the WCF tephra are unresolved, which causes
confusion on how to interpret the time series of lake level in the context of global climate. Ash 19, for instance, marks a shift to wetter conditions in the Mono Basin that persisted for nearly the entire duration of the Wilson Creek Formation. When this shift occurred is not known because there are two interpretations on Ash 19’s age. One interpretation argues in favor of a 40 ka age (Benson et al., 1990; Benson et al., 1998; Cassata et al., 2010). This 40 ka datum is based on $^{14}$C dating of lacustrine macrofossils. The alternative interpretation for Ash 19’s age, which is constrained by relative paleomagnetic intensity data, is 66 ka (Zimmerman et al., 2006).

In chapter 2 of this dissertation, I attempt to clarify the debate on the age of Ash 19. I measured the U/Th ages of two carbonate samples that were associated with Ash 19. One cross-cuts Ash 19; the other underlies it. Their respective U/Th ages are 65.6 ± 0.25 ka and 66.89 ± 2.78 ka. These dates corroborate Zimmerman et al., (2006)’s 66 ka age interpretation of Ash 19. The corroborations supports the position that the lacustrine macrofossil $^{14}$C dates are too young via modern carbon contamination. Furthermore, a 66 ka datum for Ash 19 suggests that the onset of a long-lasting wet interval in the Mono Basin corresponds with a global reorganization of climate that occurred at the transition between Marine Isotope Stages 5 and 4.

In chapter 3, I show Mono Lake's fluctuations 25-9 ka. In this study, I used stratigraphic and geomorphic evidence in conjunction with $^{14}$C and U/Th dates to constrain a time series of lake fluctuation. I measured $^{14}$C dates on bird bones and charcoal that I found in sedimentary sequences or cemented in tufa mounds. U/Th dates were measured on carbonates collected from sedimentary exposures or those outcropping on the landscape. A synthesis of the multi-pronged dataset showed that the rises and falls of Mono Lake agreed with episode of abrupt cooling and warming of North Atlantic climate. Conventional wisdom, which claims that western US wetness is tied to the position and size of the North American ice sheet, cannot explain the
variability of the lake level along with its correspondence with North Atlantic climate (e.g., Antevs, 1952). I argue that the Mono Lake data is consistent with the hypothesis that western US hydroclimate is linked with North Atlantic climate via tropical atmospheric circulation—namely, the Hadley cell (Lee et al., 2011; Chiang et al., 2014). By this hypothesis, the Hadley cell response to cooling in the North Atlantic would be to increase poleward energy transport during the cool season. An increase in energy transport would be met by an increase in the meridional delivery of subtropical and tropical storms to the western U S. Warming in the North Atlantic, on the other hand, would cause the converse: drying in the western US.

A peculiar development of this study was the fortuitous discovery of the lowest known deposit of fluvial gravels during Wilson Creek time. The deposit pointed to an extreme low stand of the lake—the lowest of Wilson Creek time. I term it the “Big Low”. In chapter 4, I present a stratigraphic analysis of strata from the Big Low type-locality as well as dispersed sites that corroborate the interpretation. The strata show that the lake fell to or below 1,975 m between the deposition of Ashes 5-4. I measured a carbonate U/Th date on a littoral conglomerate at the Big Low type-locality. And I measured U/Th dates on carbonates that were overlain by Ash 4 or that cross-cuts Ash 5. The data show that the Big Low occurred sometime between 24.3 ka – 20.5 ka. This interval encompasses the coldest interval of the last glaciation—dubbed the Last Glacial Maximum, LGM (Clark et al., 2009). Lake level records from the Searles and Lahontan Basins also show unequivocal evidence of severe drying during the LGM. The cause of these dry periods remains elusive. More precise chronologies are needed.
1.2 Enigmatic Lake Deposits in the Great Basin of the Western United States

The Great Basin of the western US is a region renown for its aridity. Playas define the majority of its valleys. Seasonal wetness grows ephemeral lakes. But few perennial lakes exist. Dryness-adapted vegetation, likewise, dominates the landscape (DeLucia and Schlesinger, 1990; Cornstock and Ehleringer, 1992). But ancient lake deposits points to the existence of a wetter climate (Figure 1.1; Russell, 1885, 1889, 1895; Gilbert, 1890). This contradiction has puzzled geologists since the late 19th century (Russell, 1885, 1889).

When lakes do not have an outlet, their surface area indicates hydroclimate (Benson and Paillet, 1989). They transgress when it gets wetter. They regress when it gets drier. A repetition of these expansions and contractions of the lake modifies the landscape. Erosion of the shorelands by wave action occurs during lake rises (Hutchinson, 1957). The integration of this erosion forms low-gradient, sublacustrine terraces. Lake falls drive inflowing streams to incise, which forms canyons. Lake falls also expose once submerged lacustrine landforms (deltas and littoral embankments, for instance). Geologists use these conspicuous features for the study of past hydroclimate.

Late 19th century geologists were the first to report on evidence for large lakes (Russell, 1885a, b; Russell, 1889, Gilbert, 1890). The studies found shorelines up to 200 m higher than the lakes or valley floors at the time. Lacustrine sediments at similar elevations corroborated their shoreline observations. The geologists argued that the present hydroclimate could not support such expansive lakes. Thus it was reasoned that past climate must have been wetter. The timing of this wet period, however, was unknown. Radiometric dating of lacustrine rocks was not workable until the mid-20th century (Libby et al., 1949; Arnold and Libby, 1949; Arnold and Libby, 1951). But geologic clues provided information on the relative timing of lake high stands. Field observations showed an association between the maximum position of ancient glaciers and high-elevation lake deposits (e.g., Russell, 1889; Putnam, 1950). The
connection between high lake level and expanded glaciers remains unresolved (e.g., Antevs, 1952; Oster et al., 2015; Lora et al., 2017).

The geologic data on the existence of large Great Basin lakes was unambiguous. But less clear was the source of the moisture that fed the lakes. The configuration for the last glaciation’s climate system is also not well resolved. In order to better constrain the dynamics of ancient climate, scientists turned to the Earth’s present climate system as analogue for the past. By using this analogy, they hypothesized on the structure of the glacial climate system.

There is uncertainty on the hydroclimatic effects of anthropogenic climate change (IPCC, 2013). General circulation models predict a reordering of global rain belts (Collins et al., 2013). Their simulations show wetter tropics and drier subtropics. In the Great Basin, model runs show reductions in both precipitation and evaporation (Seager et al., 2013). These reductions afford up to a 20% reduction in precipitation (Seager et al., 2013). Such a reduction in wetness would cause adverse effects to society. The validity of these predictions is contingent on a model’s ability to retrodict past hydroclimate. This precondition necessitates well-resolved records of natural hydroclimatic variability.

1.3 The Climate System (Wallace and Hobbs, 2006)

Solar heating is uneven on Earth. There is more at the equator. And there is less at the poles. This thermal imbalance drives energy transport from the equator to the poles, driving what is understood to be Earth’s climate system. The Earth’s rotation deflects this air flow, reordering the circulation into distinct circulation cells. There is a tropical cell, and there is a polar cell. And the boundary between the two is marked by a belt of westerly winds that is generally referred to as the “jet stream”.
Two atmospheric systems control Great Basin wetness. One is the North Pacific High. It is a high-pressure system associated with air sinking in the subtropics. Its station is the North Pacific. The other is the Aleutian Low. It is a low-pressure system centered in the North Pacific. Its position coincides with the boundary between westerly jet and the polar cell. The Aleutian Low is the source region of moisture for storms that reach the Great Basin. The path of these storms is from west to east. Thus the storms are termed the “westerlies”. The strength of these westerly storms are key elements for understanding present and past Great Basin wetness.

As a result of Earth’s tilt and orbit around the sun, the seasonal distribution of heat varies. This causes the position and strength of low- and high-pressure systems to vary. Hence wetness varies by season, too. During the cool season, the Aleutian Low strengthens and advances to the south. As a result, the North Pacific High diverts to the south. This diversion provides a corridor in which westerly storms can flow eastwards towards North America, which delivers Pacific moisture to the Great Basin. The converse occurs during the warm season. Warming weakens the Aleutian Low and strengthens the North Pacific High. This causes both systems to migrate northwards. The more northerly position of the North Pacific High blocks westerly storms. Paleoclimatologists invoke this seasonal flip-flop of atmospheric circulation to explain variations in glacial and interglacial wetness.

Cooler temperatures characterize the last glacial period (Jouzel et al., 1987). This cooling fostered the growth and persistence of an ice-sheet in North America (Dyke and Prest, 1987). The ice-sheet extended across the northern half of the continent. It is thought that this perennial cool body produced an anomalous area of sinking air, yielding a persistent, high-pressure zone about its surface (Manabe and Broccoli, 1985). Antevs (1952) suggested that the permanence of this high-pressure system perturbed atmospheric circulation. And he hypothesized that it caused the Aleutian Low to stay in a cool-season mode throughout the year. It was perennial wetness, he reasoned, which led to the rise of
lakes across the Great Basin. Following his reasoning that assigned the degree of wetness to the size of the North American ice sheet, he argued that the lake high stands occurred at glacial maximum. Likewise, he argued that lakes fell from these high stands when temperatures warmed as the glacial period waned. And finally, he suggested that lakes desiccated or shrank to their present limits in tandem with the glacial period’s end. Thus in Antevs’s model, glacial and interglacial periods are wet and dry, respectively.

At present, peak wetness at glacial maximum is conventional wisdom. Yet there is evidence to contradict this claim (e.g., Lin et al., 1998). This evidence illustrates that lake fluctuations were not positively correlated to the size of the North American ice-sheet. Rather the fluctuations matched extraordinary oscillations in North Atlantic climate (Broecker, 1994). This contradiction has caused confusion. And it has sparked debate on the mechanisms controlling Great Basin hydroclimate (Lyle et al., 2012; Lachniet et al., 2014; Oster et al., 2015). Developing well-resolved lake fluctuation records will resolve the uncertainty.

1.4 Paleo-lakes of the Great Basin

Detailed scientific study on the fluctuations of ancient Great Basin lakes began ~125 years ago (Russell, 1885). The greatest focus of these inquiries was on the region's largest ancient lakes (Figure 1.1). One is Lake Lahontan, which is in the western Great Basin (Russell, 1885; 1895). Pyramid and Walker Lakes, Nevada are extant parts of this lake. The other is Lake Bonneville (Gilbert, 1890). It existed in the northeastern corner of the Great Basin. Its surviving lake is the Great Salt Lake, Utah.

Lake Bonneville has the most completely documented sedimentary record for the interval spanning 30-10 ka (Oviatt et al., 2015). The timing of peak wetness, however, is best constrained in the Lahontan record (Adams and Wesnousky, 1998). Modelling studies rely upon these two records to infer
past hydroclimate. But neither record is ideal for hydrologic modeling. Periods of lake spillage out of the Bonneville basin prevent a quantification of past precipitation (Oviatt, 1988; McGee et al., 2012; Oviatt et al., 2015). And times of lake desiccation in the Lahontan Basin prevent an understanding of the region's past aridity (Benson and Thompson, 1987). Uncertainty as to how stream piracy affected the levels of Lake Bonneville or those of Lahontan’s sub-basin lakes further the predicament (Benson and Thompson, 1987; King, 1993; Adams, 2003). These matters make accurate hydrologic modelling a challenge.

1.5 Mono Lake and the Wilson Creek Formation

The central topic of this thesis is the fluctuations of Mono Lake, California (Figure 1.1-1.2). My study encompasses the time interval from 66 to 9 ka; however, a detailed study of the lake fluctuations encompasses 25-9 ka. This 16 kyr interval spans three distinct climatic intervals. The earliest is the glacial maximum, widely referred to as the Last Glacial Maximum. The second is the glacial period's rapid demise. And the third is the Holocene.

Mono Lake is a closed-basin lake in east-central California (Figure 1.2). It is alkaline and saline. It occupies an active tectonic depression (Pakiser, 1976; Bursik and Sieh, 1989; Rood et al., 2011). Its western margin abuts the high-relief escarpment of the Sierra Nevada. Low- to moderate-relief ranges mark its other margins.

For the purposes of hydroclimatic study, few geologic caveats characterize Mono Lake. A single moisture source, which is cool-season westerly storms (Figure 1.3), dominates its precipitation. The lake has not overflowed or desiccated during the last glacial cycle. Its tributary streams were not pirated. Nor were their waters drawn off by the basin’s sedimentary and basement rocks. A previous study quantifies slip-rates along major range-bounding faults (Rood et al., 2011). This makes elevational corrections for
tectonism straightforward. And the basin’s hypsometry suggests that its surface area varies linearly with elevation (Figure 1.4). Thus lake elevation is a robust metric of past hydroclimate.

Besides these favorable elements, the basin’s arid climate limits vegetation. This makes lucid the basin’s glacier-, stream-, and wave-formed geomorphology (Figures 1.5-1.8). The lake’s tributary streams expose up to 80-vertical meters of strata along their incised canyon walls (Lajoie, 1968). The strata comprise inter-fingering fluvial, glacio-fluvial, and lacustrine strata. The excellent exposure and numerous tephra layers are important for the development of a highly refined lake fluctuation record.

The canyon walls of the basin's tributary streams expose the Wilson Creek Formation (Figure 1.9). It is the Mono Basin’s lithostratigraphic unit of the last glaciation. And it underpins my investigation of the Mono Basin’s hydroclimatic record. Its age—and the ages of its 19 intercalated tephra—is the subject of contentious debate (Benson et al., 1998; Kent et al., 2002; Benson et al., 2003; Zimmerman et al., 2006; Vazquez and Lidzbarski, 2012). Before the thesis work presented here, there were two hypotheses on the Wilson Creek Formation age. The two models argue that the Wilson Creek Formation is 40-14 ka or 67-15 ka. The younger age-model estimate relies on the accuracy of lacustrine macrofossil $^{14}$C dates (Lund et al., 1988; Benson et al., 1998; Benson et al., 2003). The older model relies on paleomagnetic intensity correlations (Zimmerman et al., 2006; 2011). Radiometric dates on tephra that include U-Th (Vazquez and Lidzbarski, 2012) and (U-Th)/He (Cox et al., 2012) analyses on volcanic phases corroborate the estimated dates based on paleomagnetic intensity correlations.

This thesis includes three chapters based on field and geochronological data. In chapter 2, I attempt to settle the debate on the age of Ash 19, the oldest Wilson Creek Formation tephra, with a new approach. I use the U/Th method to date carbonate deposits, which underlie or cross-cut Ash 19. I show that Ash 19 must have been deposited during between $66.8 \pm 2.8$ ka and $65.4 \pm 0.3$ ka. These dates are consistent with the Zimmerman et al, (2006)’s older interpretation of Ash 19’s age, which suggests that
the \(^{14}\)C estimate is 25 kyr too young. The most parsimonious explanation for the disparity is contamination via modern atmospheric CO\(_2\). I estimate that the dated material was contaminated with ~1% modern carbon. This interpretation is consistent with evidence of modern carbon contamination from past studies (Kent et al., 2002; Hajdas et al., 2004; Zimmerman et al., 2006, 2011).

In chapters 3 and 4 of this thesis, I show evidence that challenges Antevs (1952)'s conceptual model of glacial hydroclimate. I show that the fluctuations of Mono Lake 25-9 ka do not correspond to ice-sheet size.

### 1.6 Western United States Hydroclimate and Abrupt Climate Change

Findings contrary to Antevs (1952) are not new. Broecker and Orr (1958) first noted various high lake stages occurred in the Great Basin. And it was later shown that Great Basin hydroclimate did not correspond to the ice sheet's range (Broecker and Kaufman, 1965). Variations in north European vegetation were similar in timing to Great Basin lake fluctuation (Broecker and Kaufman, 1965). This led to the speculation that the two shared a common trigger. The salt chronology of Searles Lake, California also indicated varied fluctuating wetness (Flint and Gale, 1958; Stuiver, 1964; Peng et al., 1978; Lin et al., 1998). These lines of evidence suggested Antevs's model was unfitting. But some studies have discounted the notion of multiple lake high stands. Rather they inferred that there was only one high stand of Great Basin lakes during the last glaciation (Benson, 1978; Thompson et al., 1986; Benson and Thompson, 1987; Lao and Benson, 1988; Benson et al., 1990).

Two related developments have made revisiting the timing of lake fluctuations a tractable and appealing endeavor. One is the documentation of extreme and abrupt millennial climatic events in Greenland and beyond. These events were, then, put into the context of a globally-orchestrated hydroclimate as field and analytical (\(^{14}\)C and U/Th) data became more precise.
The discovery of abrupt millennial oscillations in North Atlantic climate was seminal for paleoclimatology. Stable isotopes in a Greenland ice core showed that 21 abrupt warming events occurred during the last glacial period (Dansgaard et al., 1984). Sediment cores of the same age from the North Atlantic preserved six brief intervals of ice-rafted lithic detritus (Heinrich, 1988; Broecker et al., 1992; Bond et al., 1992; Hemming, 2004). Together the ice-core and sedimentary records showed that glacial climate was variable. It comprised millennial cooling trends (Bond et al., 1993). Brief but more extreme cooling punctuated these trends. These episodes of extreme cooling coincide with slowdowns of the Atlantic’s thermohaline circulation (e.g., McManus et al., 2004). Each cooling interval concluded with an abrupt warming. This coincided with a speed up of the thermohaline circulation. "Dansgaard-Oeschger" events are the warm episodes. "Heinrich" events correspond to the periods of ice rafting. The cooling trends are "Bond cycles".

Over the last twenty years, studies have largely affirmed the speculation of Broecker and Kaufman (1965). These studies have also confirmed the hypothesis forwarded 30 years later by Broecker (1994). Most of lakes reached their highest levels during a period of extreme North Atlantic cooling 18.6-14.6 ka (Munroe and Laabs, 2013). Paleoclimatologists refer to this period as "Heinrich Stadial 1". A more limited set of observations shows that a later high stand of lesser size occurred during the Younger Dryas, an aberrant cool period 12.9-11.5 ka (Oviatt et al., 2005; Caskey et al., 2004; Briggs et al., 2005; Adams et al., 2008). These data are evidence that lake high stands coincide with abrupt coolings of North Atlantic climate. There is no consensus on the mechanism that relays changes in North Atlantic climate to the Great Basin.
1.7 How Does North Atlantic Climate Correspond with Great Basin Wetness?

Three models explain Great Basin wetness. One is after Antevs (1952), which ties lake high stands to the North American ice sheet. Two other models hypothesize on the connection between North Atlantic and Great Basin climate (Asmerom et al., 2010; Chiang et al., 2014). Both reorient the focus from regional controls—i.e., ice sheet size—to the Earth's climate system. And both models invoke the interhemispheric temperature gradient. The gradient changes if one hemisphere cools or warms relative to the other. Extreme cooling of the of the boreal winter during Heinrich stadials is one example how the gradient can steepen. Conversely abrupt warmings during boreal winters would decrease the gradient. These paired circumstances would occur during Dansgaard-Oeschger events. The models differ in detail. Asmerom et al. (2010) argues that the mid-latitude jet shifts equatorward when the northern hemisphere winter cools and poleward during intervals of warmer northern hemisphere winters. In contrast, Chiang et al. (2014) argue that there is no significant zonal shift of the mid-latitude jet. Instead they argue that northern hemisphere winter cooling strengthens the subtropical jet. The stronger subtropical jet funnels more subtropical and tropical moisture to Great Basin lakes, which causes them to rise.

Published lake fluctuation records are ill suited to test models of millennial climate connections with western Great Basin hydroclimate. The records are not of sufficient duration to capture multiple Heinrich stadials or Dansgaard-Oeschger events. Other hydroclimatic proxies, however, do contribute additional perspectives on the debate. For instance, sea surface temperature and pollen data from coastal California show a peculiar contradiction (Lyle et al., 2012). They show that Great Basin lake high stands were concurrent with coastal dryness. These results challenge the idea that westerly storms generated in the subpolar Pacific caused lakes to rise in the Great Basin. Coastal dryness precludes such a storm track. Instead, the study concluded that the source of moisture for lake high stands was to the south. The authors speculated that the summer monsoon transported the moisture. But wetness proxy data from caves in
southwestern Arizona are taken to reject this hypothesis (Wagner, 2006; Murray, 2012). Interpretations of the Arizona cave data infer that the glacial summer monsoon is relatively weak. On the other hand, the interglacial summer monsoon is relatively strong. The present summer monsoon does not sustain lakes of the size observed in the geologic record. If the interglacial monsoon is stronger than its glacial counterpart, I would argue that the glacial summer monsoon could not be the source of moisture for Great Basin lake high stands. A different source must exist.

Three suggested sources are: tropical and subtropical moisture via the summer monsoon (Lyle et al., 2012); subpolar and subtropical moisture via the mid-latitude jet (Asmerom et al., 2010); and subtropical and tropical moisture via the winter subtropical jet (Chiang et al., 2014). Of these three possibilities, the wintertime subtropical jet hypothesis fits best with the lake fluctuation data. Storm tracks of the subtropical jet can penetrate the Great Basin from its southern limits. This southern route leaves coastal California dry, satisfying the evidence for coastal dryness. But the capability of Chiang et al., 2014’s model to explain Great Basin wetness has remained uncertain because of the limited geologic data. There are few records at that are sufficiently well-dated to ground-truth this model.

1.8 Applications of Geochronology

Until recently radiocarbon has been the conventional chronometer to date lake fluctuation records. It is a powerful method for intervals younger than 25 ka. Carbon-bearing samples (terrestrial or lacustrine macrofossils) are common in lake basins. Their analysis is routine. And the results are precise. But the 5.7 kyr half-life makes it difficult to accurately date records older than 40 ka (Godwin, 1962). Carbonate U/Th dating provides an alternative to this limitation.
Until now carbonate U/Th dating has been used rarely to date lake fluctuation records (Broecker and Kaufman, 1965; Szaby et al., 1996; Lin et al., 1996; Lin et al., 1998). The most significant obstacle for its application is significant initial \(^{230}\text{Th}\) relative to an ingrown radiogenic component, estimated by measuring \(^{232}\text{Th}\) and making assumptions about the \(^{230}\text{Th}/^{232}\text{Th}\) at the time of formation. This results in a decreasing confidence in the result (precision is worse and accuracy is difficult to evaluate). Quantifying or estimating the initial component, when it is a small enough fraction of the total, allows calculating a reliable age. Samples with negligible \(^{232}\text{Th}\) can yield ages with precisions of <1% (for instance, Lin et al., 1998).

Ideal applications of the U/Th method are known from the Bonneville basin (McGee et al., 2012; Steponaitis, 2016). In these studies, the measurements were only made on dense carbonates thought to have formed in deep waters. The majority of the samples had a negligible \(^{232}\text{Th}\). But the strategy results in an indirect measure of lake level. Here in this thesis, it is my goal to use U/Th dating in the context of the lake’s sequence stratigraphy. This approach provides a direct means to date the fluctuations of the lake.

### 1.9 Carbonate U/Th dating

The Mono Basin abounds in tufa. Physico-chemical tufa form where groundwater mixes with lake water near the shoreline. It forms massive porous structures. They range up to two meters in height. Algae colonize these free-standing sublacustrine structures. These algae precipitate dense, laminated tufa via respiration. In addition, they coat other hard sublacustrine substrates (pebbles or bedrock, for example). Tufa can also form as dense, isopachous coatings where sublacustrine springs flow through gravel. Tufa found in the sedimentary record was previously considered to be undatable (Zimmerman et al., 2006; 2012). But Xianfeng Wang (Wang et al., 2011) discovered that white, glassy, dense, and
laminated carbonate has very low $^{232}$Th/$^{238}$U. I used such samples to date the Wilson Creek Formation stratigraphy and determine a time series of lake level.

Due to the high alkalinity of the modern lake water, there are elevated concentrations of actinides (Anderson et al., 1982; Simpson 1982). This yields high hydrogenous Th. Carbonates that precipitate from these waters contain elevated initial $^{230}$Th/$^{232}$Th values. Indeed previous attempts to date Mono Basin carbonates with $^{230}$Th were acknowledged failures (see Zimmerman et al., 2006).

My collaborator Xianfeng Wang (Wang et al., 2011) conducted a pilot study to date Wilson Creek Formation tufa by the U/Th method. The study showed that high-precision dates were feasible with selective field sampling. The most reliable samples are dense, glassy and white-colored carbonate. The analytical results show low $^{232}$Th/$^{238}$U (0.05-15 ‰). With samples where $^{232}$Th/$^{238}$U is < 20, initial $^{230}$Th is negligible. The precision of the ages using the modern analytical methods developed at University of Minnesota and with the careful selection criteria were typically ~1%. Replication supports this level of uncertainty. In addition, analytical results of the most ideal samples yielded precisions as good as 0.3%.

Wang’s initial study was used to demonstrate that Mono Lake rose during Heinrich Stadial 1 and the Younger Dryas. The study also showed that the lake level fell during the Bolling-Allerod, the intervening warm period between the Younger Dryas and Heinrich Stadial 1. The data corroborated findings from Lake Lahontan. And they were also similar to those of Lake Bonneville. Rises of Mono Lake coincided with intervals of anomalous cooling in the North Atlantic. One prominent fall occurred during a period of abrupt warming. The analytical precision of the data, combined with the abundant tephra layers for internal time lines in the Wilson Creek Formation, and the extensive exposures of sediments deposited in the former shorelands of the Mono Basin, indicated the promise to develop the best-dated lake fluctuation record in the Great Basin.
Our initial results led to new questions. If cooler was wetter and warmer was drier during the deglacial interval, was the late-glacial period—which was the coolest interval of the last glacial cycle—exceptionally wet, as Antevs and others predicted? When did the lake fall from its Younger Dryas high stand? Did the lake fall during the cool interval, or instead when temperatures warmed in the North Atlantic at the start of the Holocene?

1.10 Fluctuations of Mono Lake 25-9 ka

In chapter 3, I integrate the results on the lake’s fluctuations from the last glacial maximum through the early Holocene (Figure 1.11). I used a three-pronged approach for this study: sedimentary and stratigraphic analysis of the lake’s fluctuations; geomorphic analysis; and geochronology (dominantly carbonate U/Th analyses and few terrestrial plant and animal macrofossil $^{14}$C dates). The most coherent signal I inferred from the record is that Mono Lake rose during cool North Atlantic periods, and it fell when the North Atlantic was warm. Exceptionally, the lake remained at an intermediate elevation rather than falling during the prominent warmth of the Bolling period; however, the lake did abruptly fall during the Allerod period, a subsequent period of anomalous warmth. The discovery of an extreme low stand during the glacial maximum was a second exception: for inordinate cooling corresponded with extreme drying of the Mono Basin; however, the precision of the dates constraining this low stand prevented more detailed investigation.

In chapter 4, I re-examine the fluctuations of Mono Lake during the period encompassing 25-20.5 ka using stratigraphic analysis paired with carbonate U/Th data. This interval is the coolest part of the last glacial period. And by Antevs’s hypothesis, Mono Lake should have been high. I found that this interval included both very low and very high lake levels. The most spectacularly revealed of these lake
fluctuations is marked by a gravel- and silt-filled incised canyon. The base of the canyon is marked at an elevation of 1,982 meters. This is the lowest elevation low stand of Wilson Creek time. I term it the “Big Low” (Figure 1.10). Carbonate U/Th dates indicate that the Big Low occurred sometime during the interval encompassing 24.4-20.5 ka (although it is not yet possible to define the exact date better than this, it is inferred that the “Big Low” occurred in a limited part of this extensive interval. In the interval of 24.4-20.5 ka, summer temperatures were at their lowest. Evaporation should have been at a minimum. Thus it stands to reason that—contrary to the conventional wisdom—the Mono Basin was driest at some time during peak glacial conditions, and I infer it must have been a result of a significant reduction in precipitation.

1.11 Conclusion

Debate on the age range of the Wilson Creek Formation of the Mono Basin remains active (e.g., Kent et al., 2002; Benson et al., 1990, 1998, 2003; Zimmerman et al., 2006; Vazquez and Lidzbarski, 2010; Lund et al., 2017). And an accurate age model for the section is important for global climate correlations and the paleomagenetic intensity/excursion time scale. My carbonate U/Th data constrains the depositional age of Ash 19 to be between 66.8-65.4 ka. The precision of the individual ages is excellent, and I expect future field expeditions to uncover samples which will allow better constrain the ages of all the tephra.

My carbonate U/Th data rejects the ~40 ka interpretation of Ash 19’s age, which was determined by $^{14}C$ dating of lacustrine macrofossils (Benson et al., 1990, 1998). Rather it corroborates the 66 ka interpretation of Zimmerman et al, (2006), which was based on relative paleomagnetic intensity data.
Thus the age of Ash 19, which dates the onset of wet conditions in the Mono Basin, concurs with the abrupt cooling that occurred during the transition between Marine Isotope Stages 5 and 4.

Although the late glacial fluctuations of Mono Lake are still poorly constrained in detail, the discovery of the “Big Low” and the constraint that its timing is solidly within the window of the Last Glacial Maximum (LGM) is a transformational finding for the Mono Basin. But it does not stand alone. Ages of salt deposits at Searles Lake imply several dry spells during the LGM (Lin et al., 1998). New cores from Searles Lake are likely to further clarify the timing of these dry times (Wally Broecker, personal communication). And stratigraphic evidence from the Lahontan Basin corroborate the interpretation for severe dry spells during the LGM. More lake level data in the Mono Basin to better constrain the details of its lake fluctuation during the LGM are essential for context in the Great Basin and beyond. The field and geochronology data I have acquired is the proof of concept that well-resolved lake level record in this interval is attainable with more field work and geochronology.

Paleowetness in the Great Basin of the western United States has been generally considered to be geographically uniform (Antevs, 1952; Benson et al., 1990; Spaulding, 1991). But the evidence from Mono, Searles, Chewaucan, and Bonneville Basins suggests otherwise. For example, late glacial dryness in the Mono and Searles Basins was coincident with wetness in the Bonneville Basin. This deviance of records points to distinct regional climatic zones. Defining the boundaries of these regions requires a greater spatial and temporal resolution of lake level studies. This important need for understanding the texture of hydroclimate variability necessitates more studies. Getting this right is not easy and requires extensive field and geochronological efforts. Alternatively, a failing of the studies to resolve the true lake level variability may explain the disparate records.

The deglacial record of Mono Lake fluctuation is better resolved. During deglaciation, lake rises coincide with cool intervals in the North Atlantic. And conversely lake falls concur with abrupt warmings
in the North Atlantic. The monumental fall of the lake during the early Holocene is consistent with the North Atlantic correlations established with the deglacial data. The correlation between North Atlantic and Great Basin climate warrants considering a causal mechanism that connects them.

The Bölling period, 14.7-14.1 ka, is an exception to the cool-wet/warm-dry analogy. A lake fall during the Bölling period would be most consistent with the other part of the fluctuation record. But instead, although it was on the way down from the most dramatic high stand at 16 ka, Mono Lake held a relatively high and stable position during the Bölling.

The 25-9 ka Mono Lake fluctuation record does not follow changes in the size of the North American ice sheet. The fluctuation record, therefore, is inconsistent with the predictions of the Antevs (1952) hypothesis. Instead, the record shows a strong correlation to North Atlantic climate variability. The teleconnection is best explained by a model invoking a Hadley circulation response (Lee et al., 2011; Chiang et al., 2014). This model argues that changes to the interhemispheric temperature gradient diverts the Hadley cell. This, then, modulates the strength of the wintertime subtropical jet. North Atlantic cooling strengthens the wintertime subtropical jet. The converse occurs with warming. But even though this model best explains the Mono Lake record compared to the others, I cannot unequivocally support or reject the Hadley circulation model. A few missing ingredients can resolve this uncertainty. The most essential missing ingredient is evidence of low stands coincident with Dansgaard-Oeschger warm events—finding evidence for low stands is significantly more challenging than finding evidence for high stands. Resolving these predicted decade- to century-scale events is challenging for two reasons. First, high-precision dates are necessary. This requires ideal samples—low thorium and high uranium, and these conditions are more likely to hold at high lake levels and fresher conditions. Second is that lake fluctuations tend to erode the lake’s own sedimentary record during lake rises. However, I have documented evidence of exceptional low stands in Wilson Creek strata, and expect that further field
exploration will lead to finding others. Dates on these low stands may clarify the pattern of drying in the Mono Basin. I hypothesize that these low stands correlate with warm phases of Dansgaard-Oeschger events. And the data I have acquired for this thesis show promise that more field and analytical work will allow testing this hypothesis in detail.

1.12 References:


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United States linked to rapid glacial climate shifts. *Nature Geoscience, 3*(2), pp.114-117


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Figure 1.1. Digital Elevation Model of the Southwestern US--after Broecker (2010) via Ken Adams. The limits of the Great Basin are overlain in color. The maximum surface areas of its Late Pleistocene lakes are colored in blue. And the present remnants of these giant lakes are colored in white.
Figure 1.2. NASA satellite image of the Mono Basin. The blue lines mark the streams that constitute Mono Lake’s principal inflow. These streams are Rush (1), Lee Vining (2), Post Office (3), Mill(4), and Wilson Creek(5). Boxes a and b refer to the type-locality of the Wilson Creek Formation and Big Low, respectively. The orange-colored areas mark the Mono and Inyo Craters.
Figure 1.3. NASA satellite image showing California and western Nevada. The figure here shows a cartoon of the general direction of westerly storms that contribute precipitation to Mono Lake’s hydrologic budget. Mono Lake is next to the red star.
Figure 1.4. Line plot show the relationship between elevation and predicted surface area of Mono Lake. I calculated the surface area-elevation relationship at a one-elevational-meter interval within a Geographic Information System using Digital Elevation Models after Gesch et al., (2002).
Figure 1.2. Photograph looking west towards the cross-cutting embankments that were deposited by the Bloody Canyon and Sawmill glaciers. The eastern Sierra Nevada is marked by the skyline.
Figure 1.6. Photograph looking northwest towards the Sierra Nevada. The Lundy canyon terminal moraine is in the foreground. Its outer edges are marked by linear features up to an elevation of 2,145 meters. These linear features are shorelines of Mono Lake. They formed as the lake rose to its 2,155-meter high stand. The shorelines are displaced from their original elevations by tectonism. The high watermark is disturbed by ~10-vertical meters of down-dropping of the Mono Basin along the eastern Sierra Nevada frontal fault (Rood et al, 2011).
Figure 1.7. Photograph looking northwest towards the Sierra Nevada. The brown- to black-colored hill in the foreground is Black Point. Its southern margin here has been beveled by lake rises. Its cliff-line marks the termination of a lake rise. The linear features that occur along the low-gradient surface that extends towards the lake are littoral embankments.
Figure 1.8. Photograph looking north towards the Bodie Hills and showing two stream-incised canyons. Mill Creek is on the left (west); Wilson Creek is to the right (east).
Figure 1.9. Photograph looking southwest at the western canyon walls of lower Wilson Creek. The photograph shows the sedimentary sequence that defines the Wilson Creek Formation and its five-part ash sequence: Ashes 19-18; Ashes 17-16; Ashes 15-8; Ashes 7-5; and Ashes 4-1. The prominent dark layer represents the basaltic Ash 2. Two age models are debated for the Wilson Creek Formation: one argues that it encompasses 15-40 ka (Lund et al., 1988; Benson et al., 1998; Benson et al., 2003); and the other argues that it encompasses 15-66 ka (Kent et al., 2002; Benson et al., 2003; Zimmerman et al., 2006; Cox et al., 2012; Vazquez and Lidzbarski, 2012). Chapter 2 of this dissertation informs on the disagreement on the interpretation for the age of Ash 19.
Figure 1.10. Photograph looking west at the eastern canyon wall of Mill Creek. The image shows the Big Low type-locality, which is defined by fluvial and littoral gravels (G1) that overlie lacustrine silts that contain Ashes 15-8. Lacustrine silts with Ashes 7-5 are found in G1 as a slump block. Fluvial gravels (G2) cross-cut both G1 and S1. And lacustrine silts containing Ashes 4-1 (S2) overlie G1 and G2. The sequence can be interpreted to represent two periods of lake fall with concomitant fluvial incision. The fluvial disconformities for these lake falls are represented by the yellow- and red-dashed lines. The rises from the final low stand is represented by the lacustrine silts that comprise S2. These silts represent strata that filled in the ancient canyon that was incised by a stream during the second lake fall. Together these evidence of lake falls constitute what I refer to as the “Big Low”. The basal elevation of the G1 and G2 fluvial gravels is 1,982 m. This elevation is a maximum elevation for the Big Low.
Figure 1.11. My interpreted Mono Lake hydrograph for ~25-9 ka. The blue data are U/Th and \(^{14}C\) dates from this dissertation. Data points “a” and “b” refer to the corrected U/Th and uncorrected U/Th (maximum age) dates on calcite that cement the littoral embankment associated with the Big Low. The lake level curve shows a rise from the Big Low by ~20.5 ka and a rise to its 2,155-meter high stand by ~16 ka. Two lake falls occur during the period encompassing 16-14 ka. A second monumental lake rise occurs ~13-12 ka to 2,089 m. This is followed by a lake fall to ~1,980 by 11 ka. The lake then fluctuated between at least the elevations of 1,965 and 1,975 m from 11-9.5 ka. The lake rises coincide with periods of North Atlantic cooling, which are highlighted in the blue-dashed lines (Heinrich Stadial 2, 24.5-23.5; Heinrich Stadial 1, 18.6-14.7; and Younger Dryas, 12.9-11.7 ka). The lake, too, fell largely during periods of warmth (Bølling-Allerød, 14.7-12.9 ka; and the Holocene, which began ~11.7 ka). The age control on the Big Low is insufficient to determine precisely when it occurred, but it must have occurred at some time.
2.1 Abstract

The last glaciation corresponded with wetter hydroclimate in the Great Basin of the western United States. The primary evidence for this is lake fluctuation records in hydrologically closed basins. Attempts to understand the relationship between the higher lake levels and global climatic patterns is challenged by poorly-resolved age control and fragmented sedimentary records. The onset of relatively high lake level conditions during the last glaciation is well-exposed in the Mono Basin, California. Its Late Pleistocene lithostratigraphic unit, the Wilson Creek Formation, contains a tephra, named Ash 19. Its deposition marks the onset of Mono Lake’s rise during the last glaciation. But two disparate interpretations on the Ash’s age—40 ka (Benson et al., 1998) and 66 ka (Zimmerman et al., 2006)—cause uncertainty on the timing of the lake’s rise. Here I measured U/Th dates on carbonates that cross-cut and underlie Ash 19. Their respective dates, which are minimum and maximum constraints on the depositional age of the Ash, are 65.6 ± 0.25 ka and 66.89 ± 2.78 ka. These data corroborate the 66 ka interpretation of the Ash’s age. And it highlights that the start of wet conditions during the last glaciation coincided with the dawning of Marine Isotope Stage 4, a time of monumental restructuring of Earth’s climate system that included a rapid drawdown of atmospheric CO₂. The coincidence between wetter hydroclimate and lower atmospheric CO₂ is suggestive of a causal link.
2.2 Introduction

It is well agreed upon that it was wetter than present during the last glacial period in the Great Basin of the western United States. The evidence supporting this claim predominantly derives from the study of ancient lakes in hydrologically closed basins. The earliest of such studies showed evidence for lake levels that greatly exceeded the present lakes (Russell, 1885, 1889, 1895; Gilbert, 1890). And because of the lakes’ hydrologic closure, the elevated levels were interpreted to indicate a wetter climate. The timing of the high lake stands was first determined from their association with morainal embankments formed during the last glacial cycle (Russell, 1889). This relative chronology was put into an absolute context following the advent of the radiocarbon dating method (Arnold and Libby, 1949; Libby et al., 1949). $^{14}$C dates from the high lake stand deposits showed ages that concurred with the last glaciation, which is now known to cover the time 110-15 ka (Broecker and Orr, 1958; Anderson et al., 2004). Hence the general understanding that the last glacial period was wetter than the historical era.

In spite of the evidence pointing to the existence of wetter conditions during the last glacial period, there remain questions as to when this epoch of wetness began and how long it persisted. The onset and duration of glacial wetness are important data that can test the validity of hypotheses concerning the dynamics of Earth’s paleo-hydroclimate. The prevailing model, which is based on lake fluctuation data, argues that Great Basin wetness concurred with the maximum extent and size of the North American ice sheet, ~30-20 ka (Antevs, 1952, Bartlein et al., 1998). In the context of this hypothesis, periods of high lake levels prior to 30 ka cannot be explained. But there are too few observations older than 30 ka to confirm or reject the model. This obstacle is a reflection of two problems.
The first problem concerns the fragmented nature of the sedimentary record. While there is evidence showing that the Bonneville phase of Utah’s mega-paleolake was relatively high 30-15 ka, the ages of sedimentary deposits from older lake phases are debated (Nichizawa et al., 2013; Oviatt et al., 2014; Nichizawa et al., 2014; Oviatt et al., 2015). And the effects of river diversion into the lake basin ~55 ka make it difficult how to parse out the relative effects of climate on the lake levels (Pederson et al., 2016). Missing time in the sedimentary record by desiccation further adds to the confusion. For example, there is evidence showing relatively high levels of Lake Manix, which is located in the Mojave Desert, as old as 45 ka (Reheis et al., 2015). But repeated desiccations of Lake Manix preclude a measure of the hydroclimate during that time (Reheis et al., 2015). Continuous records are the remedy to this problem.

The second problem that inhibits our understanding of glacial wetness concerns chronology. The majority of the data constrain lake levels that are younger than 30 ka (e.g., Benson et al., 2013; Oviatt et al., 2015). But there are some notable exceptions that are as old as 45 ka (Lake Manix, for instance; Reheis et al., 2015). Beyond this time, the data is debated. This chronologic curtain reflects the theoretical and practical limitations of the $^{14}$C method. Because of the 5,730-yr half-life of $^{14}$C, the radiocarbon method cannot resolve ages older than 60-50 ka (Godwin, 1962). But the more practical limit is ~45 ka: for material older than 45 ka is sufficiently low in radiocarbon that it is highly susceptible to the effects of modern carbon contamination via atmospheric CO$_2$. Step-leaching experiments demonstrate that modern carbon contamination can significantly perturb the apparent radiocarbon age (Kent et al., 2002; Hajdas et al., 2004; Zimmerman et al., 2012). Only a contamination of 1% is needed to shift a radiocarbon dead sample (say, > 60 ka) to an age of 40 ka. Thus the research studies that use the $^{14}$C method are poorly equipped to date the lake histories during the last glacial cycle.
Here I re-evaluate the timing of the initial rise of Mono Lake, a closed lake in east-central California, during its most recent high-lake-level phase of the last glaciation (Figure 2.1). Evidence of this period of high lake level is recorded by sediments that compose the Wilson Creek Formation, the Mono Basin’s Late Pleistocene lithostratigraphic unit (Figures 2.2-2.3; Lajoie, 1968; Zimmerman et al., 2006). At its type locality, which is located along the lower stretches of Wilson Creek, the Wilson Creek Formation comprises 16 m of lacustrine silts and clays that are intercalated with 19 tephra (Figures 2.2-2.4; Lajoie, 1968). At its type locality, the Wilson Creek Formation is underlain by fluvial cobble gravel. And because no subaerial deposits were recognized within the Wilson Creek Formation type section, it was inferred that the level of Mono Lake had always remained above the type section’s elevation, ~1,990 m—nine meters higher than the lake’s elevational range during the last 4 kyr (Stine, 1990). The relative time for the onset of this epoch of wetness is defined by Ash 19. At the type locality, Ash 19 is intercalated in silts that are immediately underlain by the pre-Wilson Creek Formation fluvial gravel (Lajoie, 1968). The contact between the silts and the gravels represent a flooding surface, which marks the initiation of Wilson Creek time and the epoch of wetness that it concurs with.

Although the relative age of the lake’s rise can be approximated to the deposition of Ash 19, the absolute age of Ash 19 is debated (Benson et al., 1998; Kent et al., 2002; Benson et al., 2003; Zimmerman et al., 2006; Cassata et al., 2010; Vazquez and Lidzbarski, 2012). There are two end-member hypotheses for the age of Ash 19. One hypothesis, herein termed the Benson model, proposes that Ash 19 is 40 ka (Benson et al., 1998; Benson et al., 2003). This datum was determined by a linear interpolation of $^{14}$C dates on ostracod shells, tufa, and carbonate nodules (Benson et al., 1998; Benson et al., 2003). The other hypothesis is underpinned by relative paleomagnetic intensity correlations to the global paleomagnetic intensity stack (Zimmerman et al., 2006). This hypothesis, herein termed the
Zimmerman model, infers an age of 66 ka for Ash 19. A third interpretation of Ash 19’s age uses the Benson model’s underpinning $^{14}$C data to constrain a relative paleomagnetic intensity timescale for Ash 19 (Cassata et al., 2010). This effort yielded two interpretations: one that estimated Ash 19’s age to be 40 ka, which agreed with the Benson model’s interpretation; and one that suggested that Ash 19’s age was ~70 ka, concurring with the Zimmerman model’s 66 ka interpretation.

In this study, I attempt to clarify the debate on the age of Ash 19. I use the U/Th method to date carbonates that underlie and cross-cut Ash 19. The underlying and cross-cutting carbonate samples were sampled from stream-cut sedimentary exposures along lower Wilson Creek and Bridgeport Creek, respectively. I measured the sedimentary sequences the carbonates occur in, and I identified the tephra contained in the sequences following previous descriptions of the localities by Lajoie (1968)—Lajoie’s IV-E and IV-D along Bridgeport Creek (Figures 2.3 and 2.6), and Lajoie’s II-B at lower Wilson Creek (Figure 2.4). U/Th analyses of the carbonate samples provide closely-bracketing, maximum and minimum constraints on Ash 19’s age (Table 2.1). From these U/Th data, I conclude that the Ash is $67.5 \pm 2.2$ ka.

2.3 Sedimentary and Stratigraphic Context for Carbonate U/Th Data on Ash 19

2.3.1 Ash 19 along lower Wilson Creek. In the sediments exposed along lower Wilson Creek, identification of Ash 19 is unambiguous: for it lies near the base of lacustrine silts that are underlain by fluvial gravels (Figure 2.4; Lajoie, 1968). This stratigraphic relationship is easily traced along the entire exposure of the basal Wilson Creek Formation.

At one exposure along the eastern canyon wall of lower Wilson Creek, Ash 19 is underlain by a ~0.4-meter-tall tufa mound. Colluvium obscured the detail of the sediments that underlie the tufa mound. Nearby exposures reveal the typical association of fluvial gravels below the stratigraphic level of the tufa.
mound. Friable, massive, grey-colored calcite compose the tufa’s structure. The texture and structure of the tufa is consistent with its formation as a sublacustrine spring. The silts that contain Ash 19 suggest a low-energy lacustrine depositional environment. The stratigraphic relations, therefore, suggest that Ash 19 was deposited into the lake after the formation of a sublacustrine tufa mound. I sampled a scalenohedral calcite crystal from the tufa’s interior (Figure 2.5) for U/Th dating. Because the tufa mound formed before the deposition of Ash 19, a date on the tufa mound constrains the maximum depositional age of Ash 19. And because the tufa mound is directly overlain by the Ash, it is a very close constraint on the depositional age, albeit not a direct constraint.

2.3.2 Ash 19 along Bridgeport Creek. Lajoie measured five sections (B-G) of the sediments exposed along Bridgeport Creek canyon (Figures 2.1, 2.3, and 2.6). By Lajoie’s reporting, two of the sedimentary sequences, IV-D and IV-E, contain Ash 19. In both sequences, Ash 19 is intercalated in silts that are immediately underlain by alluvial strata. This is similar to the stratigraphic context in which Ash 19 was deposited in the Wilson Creek Formation type section. But what differs at Bridgeport Creek is that the sedimentary sequences overlying Ash 19 are incomplete (i.e., missing tephra). A simple counting scheme, therefore, cannot be used to identify unknown tephra, leaving the identification of non-unique tephra uncertain. It is because of this uncertainty, I reason, that Lajoie’s tephra correlation of Ash 19 was tentative; however, allanite U/Th data corroborate Lajoie’s interpretation that the basal tephra in IV-E is Ash 19 (Ruprecht et al., in prep). The tephra identification matters because I sampled and dated a carbonate deposit that cross-cut Lajoie’s Ash 19 in IV-D. This datum yields a minimum depositional age for Ash 19. Here I re-measure a part of IV-D to provide additional context for
the carbonate that cross-cuts Ash 19. I only measured a part of the sequence because I wanted to refrain from encroaching onto private land. As a result of the fragmented sequence, I could not directly relate my observations to those of Lajoie (1968). Thus I also re-measure a part of IV-E to identify Ash 19’s characteristics and depositional context. I use these data from IV-E to corroborate my interpretations for ash identification in IV-D.

2.3.2.1 Lajoie (1968)’s IV-E. My re-measuring of Lajoie (1968)’s IV-E shows a two-part sequence (Figures 2.1, 2.3, and 2.7). The lower half of the measured section contains two fining-upward sequences. The lowest sequence comprises 58 cm of poorly sorted cobble gravel that is overlain by 48 cm of medium to very coarse cross-laminated sands. The upper sequence overlies this bipartite sequence. Its base is marked by an erosional contact. And the erosional contact is successively overlain by: 1) 10.5 cm of white, cross-laminated tephra that abuts and overlies a horizon of interspersed boulders (Figure 2.8); and 2) 36 cm of thinly-laminated silts, clayey silts, and one intercalated pink, three-mm-thick tephra. The silts coarsen upward to unconsolidated, very coarse sands. This interval of coarsening-upward strata is 336.5-cm-thick. And it is overlain by two successive fining-upward sequence. The lower sequence is 30-cm-thick; the upper sequence is 40-cm-thick. The 40-cm-thick sequence contains a white rhyolitic tephra, which was interpreted by Lajoie (1968) as Ash 15. The basal contact of each of these sequences is erosional, and it is marked by a one-cm-thick granule to pebble gravel lag. The basal elevation of the lower sequence’s gravel is traceable at the same elevation, ~2,019 m, for at least ~100 m.

I interpret the lower fining-upward sequence of gravels and sands as fluvial deposits. I interpret the gravel of the second fining-upward sequence as fluvial gravel, and the basal contact to be an erosional disconformity carved by stream incision. It is conceivable that the silts and clayey silts that overlie the fluvial
gravels are also fluvial deposits. In either case, they reflect deposition into a low-energy environment. The silts’ extensive lateral continuity—both down- and up-stream—make more plausible the interpretation that the silts are lacustrine deposits. Accordingly, the vertical succession of fluvial to lacustrine strata indicates lake flooding of a subaerial environment. Furthermore, I argue that the silt’s two intercalated tephra were deposited in the lake.

The coarsening upward sequence, which grades from very fine to very coarse sands, could have formed by either mass wasting of an inset delta front (for terminology see Stine, 1987), or by sands swept by littoral currents across lake-flooded interfluvial surfaces and deposited into lake. For the purposes of this study, it is only relevant that both scenarios would lead to the conclusion that the strata are sublacustrine deposits.

For the gravel layer overlying the silts, I reason that the erosional surface underlying the gravels is likely a product of wave planation. Wave action bevels strata along the shore zone, and as a lake rises, this beveling will continue range-ward, forming a low-gradient erosional surface (Hutchinson, 1957). Those sediment clasts too large to be transported by the littoral currents remain on the wave-planed surfaces as lag deposits. Thus each disconformity-gravel-silt package records a lake lowering followed by a rise. In the section described here, there is evidence for three lake level falls and rises, with the uppermost silts containing Ash 15 representing the final rise. Lajoie’s IV-E continues above Ash 15, but it was not measured here to avoid trespassing onto private land.

Allanite U/Th data suggest that the basal tephra is Ash 19. Thus if Lajoie’s interpretation that the upper tephra is Ash 15, the tephra that lies between Ash 15 and Ash 19 must be one of these three tephra: Ash 18, Ash 16, or Ash 17. Two of these three are missing due to episodes of erosion, which are likely highlighted by the wave-cut surfaces described above. But for the questions asked here, the unknown tephra’s identification is not relevant.
2.3.2.2 Lajoie’s IV-D. Carbonate cross-cuts the basal tephra in IV-D, so I re-measured a part of this section (Figure 2.9). I referenced my observations at IV-E to help identify the unknown tephra. I measured an ~3.4-meter-thick fining-upward sequence from the base of the eastern canyon wall of Bridgeport Creek (Figures 2.1 and 2.3). From the base of this sequence to ~0.5 m, the section is obscured by vegetation. Above this interval is a 2.82-meter-thick (G1) interval of poorly sorted gravel with clasts up to boulders in size. This gravel is overlain by a ~10-cm-thick deposit of thinly laminated, white, rhyolitic tephra (T1, Figure 2.10) that is cross-cut by stacks of planar-parallel to botriodal, white to pale-yellow carbonate deposits. Each depositional unit of the carbonate resembles a deposit from a cave (i.e., they grow down from roofs and up from floors). And each of the deposits is laminated. The carbonates are overlain by a sandy pebble horizon, which is overlain by silt (S1, Figure 2.9). Details of the silt’s structure were obscured by vegetation.

I interpret the poorly sorted boulder gravel as fluvial deposits. And I infer that the tephra observed in this sequence is the same as the basal tephra--Ash 19--in IV-E. Their matching thickness, color, and stratigraphic position at the base of the lacustrine silts supports this interpretation. I sampled the carbonate for U/Th dating to provide a minimum depositional age for Ash 19.

2.4 U/Th Age Estimates of Ash 19

The U/Th analyses were conducted at the Earth Observatory of Singapore, Nanyang Technological University. The carbonates samples from Bridgeport and lower Wilson Creeks were selected for high density, white color, and an apparent absence of detritus. These samples were cleaned by sonication in ultra-
pure water and later air dried. Once dry, the sample from Bridgeport Creek was powdered using a carbide dental tool. I milled 1 mg of sample. But powdering of the lower Wilson Creek sample, which was a single scalenohedral calcite crystal, was not possible because it fractured under the pressure of the dental tool. Thus as an alternative to powdering, I used a fragment of the scalenohedral crystal for the sample’s analysis. Following this initial preparatory step, the samples’ U and Th isotopes were separated in a metal-free chemistry clean room following a methodology similar to Edwards et al., (1986, and 1987) and Cheng et al., (2000). The U and Th sample fractions were, then, measured using a ThermoFisher Neptune Plus multi-collector ICP-MS. The samples’ ages were calculated using the $^{230}$Th ($\lambda_{230} = 9.1705 \times 10^{-6}$ yr$^{-1}$) and $^{234}$U ($\lambda_{234} = 2.82206 \times 10^{-6}$ yr$^{-1}$) half-lives of Cheng et al., (2013) and the $^{238}$U ($\lambda_{238} = 1.55125 \times 10^{-10}$ yr$^{-1}$) half-life of Jaffey et al., (1971). The detrital/hydrogenous $^{230}$Th correction for age calculations assumes an initial atomic ratio of 10 ($\pm 5$) $\times 10^{-6}$, which is significantly greater than what would be assumed for material at secular equilibrium with the bulk earth $^{232}$Th/$^{238}$U ratio of 3.8. This $^{230}$Th correction is based on the evidence for elevated actinide concentrations in the modern lake water (Anderson et al., 1982). And the error in the initial ratio is arbitrarily assumed to be 50%.

The U/Th date on the carbonate from lower Wilson Creek, which was overlain by Ash 19, is 66.89 ± 2.78 ka (Table 2.1, Figure 2.5). And the U/Th date on the carbonate from Bridgeport Creek, which cuts across Ash 19, is 65.6 ± 0.25 ka (Table 2.1, Figure 2.10). Therefore the deposition of Ash 19 is constrained by these observations and dates to have occurred between 66.89 ± 2.78 ka and 65.6 ± 0.25 ka.

2.5 Discussion

The ~66 ka carbonate U/Th age for Ash 19 corroborates the Zimmerman model, which is generally consistent with the age interpretations from U-Th dating of the tephra layers (Vazquez and Lidzbarski,
2012). On the other hand, it appears that the Benson model estimate for Ash 19 is approximately 25 kyr too young. I argue that the Benson model’s anomalously young age interpretation is a reflection of the limit of reliable radiocarbon measurements, which is far exceeded by the age of the sample. The simplest explanation for the discrepancy is modern carbon contamination, as suggested by Kent et al. (2002).

The carbonate U/Th constraint on Ash 19 agrees with the older of two interpretations, ~70 ka, forwarded by Cassata et al., (2010). But because Cassata et al., (2010)’s relative paleomagnetic intensity correlation is entirely dependent on $^{14}$C dates, which are demonstrably affected by modern carbon contamination and perhaps an unconstrained reservoir effect too, I argue that prudence needs to be exercised before validating the study’s interpretation. While I acknowledge the study’s efforts and results, I favor the Zimmerman model’s methodology and, therefore, its results and interpretations.

Two $^{14}$C measurements on tufa at the stratigraphic level of Ash 19 afford an estimation of the magnitude of young carbon contamination (Benson et al., 1990). They yield ages of ~43 ka and ~32 ka (Benson et al., 1990). Because these samples must be ~66 ka—and, therefore, radiocarbon dead—the radiocarbon activity of these ages imply a contamination of ~1% and ~3% young carbon, respectively. It was reported by Benson et al. (1990) that younger carbonate coated the 32 ka sample. This suggests that its young age is partly due to secondary carbonates precipitated from groundwater. The 43 ka sample was not reported to be coated by secondary carbonate. Thus its young age may be entirely due to adsorbed modern atmospheric CO$_2$, which is argued to be unavoidable on any air-exposed surfaces (Paul et al., 2016). Acid-leaching experiments corroborate this interpretation by demonstrating the presence of modern carbon in ostracod shells and other carbonates of Wilson Creek age (Kent et al., 2002; Hajdas et al., 2004; Zimmerman et al. 2012). Thus I reason that contamination levels of ~1% are a reasonable estimate of contamination expected in other carbonate samples in the Wilson Creek Formation.
Published data indicates a 61.7 ± 1.9 ka allanite and zircon U-Th date on the first coarse tephra below Ash 17 at South Shore. This tephra was inferred to be Ash 19 (Vazquez and Lidzbarski, 2012). But a 61.7 ± 1.9 ka datum of Ash 19 is not quite consistent with the 67.5 ± 2.2 ka carbonate U/Th estimates of Ash 19 reported here. New allanite U/Th analyses from tephras collected from South Shore and Wilson Creek Canyon suggest that the 61 ka date is, in fact, analyzed from a tephra that is younger than Ash 19 in the South Shore section (Ruprecht et al., in prep). The new allanite U/Th analyses from Ash 19 at the Wilson Creek Formation type-section agree with the 66 ka estimates from Zimmerman et al. (2006) and the carbonate U-Th data reported here (Ruprecht et al., in prep).

The age of Ash 19 and the base of the lacustrine silts of the Wilson Creek Formation in the Mono Basin is 67.5 ± 2.2 ka. Thus Mono Lake rose to levels that characterized nearly all of the remaining part of the Wilson Creek Formation at approximately this time. This hydroclimatic pivot occurs during Marine Isotope Stage 4 (74-59 ka, Martinson et al., 1987). Marine Isotope Stage 4 corresponds with a drawdown of atmospheric CO$_2$ from ~250-205 ppm that took place 69-64 ka (Petit et al., 1999; Ahn and Brook, 2008).

And for the remaining 50 kyr of the last glaciation, ~69-19 ka, atmospheric CO$_2$ remained below 220 pm. The coincidence between this 50 kyr interval of depressed atmospheric CO$_2$ and the relatively high lake levels of the Wilson Creek Formation (~68-15 ka) is suggestive of a causal link; however, the causal factors of the wetter hydroclimate may not be directly related to the global cooling brought on by the fall in atmospheric CO$_2$. But the relatively high lake levels may be influenced by other agents that were impacted by the drop in CO$_2$ (changes in the position or strength of the westerly storms, for instance).
2.6 Conclusion

The age of Ash 19 is 66.9 ± 2.8 ka. This datum constrains the time when Mono Lake rose, and it marks the beginning of the Wilson Creek Formation to be coincident with Marine Isotope Stage 4. The evidence presented here shows that the lake fluctuated throughout the last glacial cycle. And during the times of lake transgressions, wave planation removed some of the Wilson Creek Formation deposits, including two of the following three tephra at Bridgeport Creek: Ashes 18-16. Furthermore, the observations and data reported here suggest the possibility of further refinement of the lake fluctuation history throughout this interval.

2.7 References


Benson, L.V., Lund, S.P., Burdett, J.W., Kashgarian, M., Rose, T.P., Smoot, J.P. and Schwartz,


Oviatt, C.G., 2015. Chronology of Lake Bonneville, 30,000 to 10,000 yr BP. *Quaternary Science Reviews*, 110, pp.166-171.


Ruprecht et al., (2017) Allanite U-Th ages of Wilson Creek Formation tephra. Manuscript in


Table 2.1: Carbonate U/Th data

Note: Corrected 230Th ages assume a 230Th/232Th atomic ratio of 10 ± 5 x 10^-6, which is similar to the present lake waters (Anderson et al., 1982).

<table>
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<th>Sample ID</th>
<th>Location</th>
<th>Context</th>
<th>238U (ppb)</th>
<th>232Th (ppt)</th>
<th>234U measured</th>
<th>a ([230Th/238U] activity)</th>
<th>b</th>
<th>c</th>
<th>Age (kyr) uncorrected</th>
<th>Age (kyr) corrected b,c</th>
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<td>S553</td>
<td>Bridgeport Creek</td>
<td>cross-cuts Ash 19 (minimum depositional age)</td>
<td>2637 ± 4</td>
<td>1065 ± 26</td>
<td>308.3 ± 2.3</td>
<td>0.6046 ± 0.0013</td>
<td>65718 ± 245</td>
<td>65699 ± 245</td>
<td>371 ± 3</td>
<td>65633 ± 245</td>
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<tr>
<td>GA-19</td>
<td>Wilson Creek</td>
<td>underlies Ash 19 (maximum depositional age)</td>
<td>7324 ± 10</td>
<td>487327 ± 9766</td>
<td>120.7 ± 2.4</td>
<td>0.5410 ± 0.0012</td>
<td>70859 ± 312</td>
<td>66592 ± 312</td>
<td>146 ± 3</td>
<td>66887 ± 312</td>
</tr>
</tbody>
</table>

B.P. stands for "Before Present." Present is defined as 1950 A.D.
Figure 2.1. NASA satellite image showing the Mono Basin and sites of interest for this study. The blue lineaments highlight Mono Lake’s principal inflowing streams: (1) Rush; (2) Lee Vining; (3) Post Office; (4) Mill; (5) Wilson Creek. The red stars indicate the two localities—lower Wilson Creek and Bridgeport Creek, from west to east—where I collected and dated carbonates associated with Ash 19.
Figure 2.2. Photograph looking southwest at the western canyon wall of lower Wilson Creek. The sedimentary sequence that is exposed represents the Wilson Creek Formation type section. The colored horizons show the designated stratigraphic positions of various ashes (Ashes 4-1 at the top, for instance).
Figure 2.3. A modified version of a general cross-section of the Wilson Creek Formation stratigraphy from Zimmerman et al., (2011). The figure shows Lajoie (1968)’s measured sections IV-E and IV-D from Bridgeport Creek canyon. IV-D and IV-E along with the lower Wilson Creek type section locality are transposed onto the cross-section to show how they relate in the context of the basin’s stratigraphy. Lettered ash sequences A-E refer to the following groups of tephra: Sequence A, Ashes 1-4; Sequence B, Ashes 5-7; Sequence C, Ashes 8-15; Sequence D, Ashes 16-17; Sequence E, Ashes 18-19.
Figure 2.4. Lajoie’s stratigraphic interpretation—from Lajoie (1968)—of the sediments exposed in the lower Wilson Creek canyon, which constitute the type section of the Wilson Creek Formation.
Figure 2.5. Photograph of the tufa sample, GA-19, that is overlain by Ash 19 in the canyon walls of lower Wilson Creek canyon (see section II-B in Figure 2.4). U/Th analysis of a scalendohedral calcite resulted in a datum of ~67 ka (Table 2.1). This datum represents the maximum depositional age of Ash 19.
Figure 2.6. Lajoie’s stratigraphic interpretation—from Lajoie (1968)—of the sediments exposed in the Bridgeport Creek canyon. The two sites I revisited in this study are IV-E and IV-D.
Figure 2.7. My measurement of section IV-E of Lajoie (1968) at Bridgeport Creek (see Figures 2.1 and 2.6). The identification of the second-lowest tephra as Ash 18 is speculative. This tephra could also be Ashes 16 or 17. Layers highlighted in red are tephra.
Figure 2.8. Photograph from Lajoie (1968)’s IV-E showing cobble gravel overlain by Ash 19, which is in turn overlain by lacustrine silt. The sedimentary relations are very similar to the basal tephra in section IV-D.
Figure 2.9. My measurement of section IV-D of Lajoie (1968) at Bridgeport Creek (see Figures 2.1 and 2.6). The layer highlighted in red is a tephra.
Figure 2.10. Photographs showing the stratigraphic context of the carbonate sample (S553, Table 2.1 for data) at section IV-D. A) Sampling location. G1-cobble gravels. T1-tephra inferred to be Ash 19. S1-lacustrine silt. B) zoomed in view of sample S553.
Chapter 3: Late glacial and deglacial fluctuations of Mono Lake, California

3.1 Abstract

Published closed lake fluctuation records from the Great Basin of the western US show that late glacial and deglacial hydroclimate was highly variable. Using these lake level records to inform regional paleoclimate is, however, challenging: for the temporal resolution of the physical expressions of lake level is limited. This problem is compounded by spotty sedimentary exposures and difficulty with dating. Accordingly, these limitations curb our understanding of the mechanisms triggering the changes in hydroclimate. Here I present a U/Th- and \(^{14}\)C-dated sedimentary and geomorphic record from the Mono Basin, a closed-lake basin in east-central California, for the time encompassing 25-9 ka. My data show that lake level fluctuations follow the pattern of North Atlantic climate variability. Relative lake high stands coincided with North Atlantic cooling. Conversely, relative lake low stands coincided with North Atlantic warming. One apparent exception to this coincidence of millennial variability concerns a late glacial low stand. Although the precise timing is not resolved, the best estimate by direct dating of this low stand’s age is 22 ± 2 ka, and it can be confidently placed within the time interval of <24 ka and >20.5 ka. Taken together, the lake fluctuations also show a coincidence with records from subtropical and tropical hydroclimatic proxies. Such harmony is most consistent with the hypothesis that global rain belts are influenced by North Atlantic climate via Hadley circulation. By this hypothesis, transgressions and regressions of Mono Lake are manifestations of a strengthening and weakening of the wintertime subtropical jet owing to northern hemisphere thermal forcing.
3.2 Introduction

The Great Basin of the western US is renown for its numerous hydrologically closed lake basins. In principle, the water balance of these lakes is simple (Benson and Paillet, 1989). A lake expands if inputs from direct precipitation and inflow exceeds evaporative demand. And it contracts when evaporative demand exceeds these inputs. Thus the lake’s fluctuations are a measure of hydroclimatic change. These fluctuations, however, can be more nuanced in large lakes with sub-basins (e.g., Lake Lahontan; Russell, 1885).

Modelling studies predict that the warming due to anthropogenic emissions will cause extreme droughts across the Great Basin (e.g., Seager et al., 2012; Cook et al., 2015). The projected reductions in long-term water availability will have significant negative societal consequences. But the instrumental record's time series is too limited a time span to fully test the range of possible variability. The geologic record is thus needed to meet the shortfall. Records of ancient lake variability provide an opportunity for ground-truthing needed to test model predictions. Such ground-truthing allows greater confidence in simulations of future climate.

Since the late 19th century, it has been well known that vast lakes once occupied the Great Basin (Russell, 1885, 1895; Gilbert, 1890), implying that the Great Basin was significantly wetter than the present sometime in the past. But the mechanism controlling the shift from wet to dry has been debated (Antevs, 1952; Kutzbach and Wright, 1985; Benson et al., 1990; Bartlein et al, 1998; Benson et al., 1998; Benson et al., 2003; Lyle et al., 2012; Munroe and Laabs, 2012; Benson et al., 2013; Oster et al., 2015; Lora et al., 2016; Wong et al., 2016; Lora et al., 2017). Resolving this debate requires well-resolved and well-dated records.
Here I present a carefully documented record of lake fluctuation from the Mono Basin, California. The record encompasses 25-9 ka. I used carbonate U/Th and terrestrial macrofossil $^{14}$C dates to date the fluctuations of the lake. My data show that Mono Lake fluctuated in tandem with abrupt changes in North Atlantic climate during the late glacial, deglacial, and early Holocene.

Mono Lake is a perennial body of saline and alkaline water. It rests in an 1,800-square-kilometer, hydrologically closed basin in east-central California (Figure 3.1). The 3,800-meter-high Sierra Nevada marks the basin’s western margin. The lake's principal inflow comprises runoff derived from Sierran snowmelt, which derives from precipitation by way of cool season westerly storms (Voster, 1985; Serreze et al., 1999).

At present, Mono Lake’s level is 1,945 m. The lake fluctuated over 40-vertical meters during the last 4 kyr (Stine, 1990). During this 4-kyr interval, the shoreline rose as high as 1,981 m. And it twice regressed to levels as low as 1,941 m. Little is known about the lake’s fluctuations between 4 ka - 11.5 ka (Benson et al, 1990). But it is well understood that the lake’s Late Pleistocene elevations were much higher and also of greater range (Russell, 1889; Lajoie, 1968). The high stand of the last glacial cycle reached 2,155 m (Russell, 1889), and as shown in this thesis, the lowest-documented stand of the last glacial cycle reached an elevation between ~1,975-1,955 m, or perhaps a few meters lower than this (see Chapter 4, this dissertation).

Road-, wave-, and stream-cuts expose Mono Basin's sedimentary deposits. The basin’s Late Pleistocene lithostratigraphic unit is the Wilson Creek Formation (first named and described by Lajoie, 1968). It comprises glacial, littoral, fluvial, and lacustrine sediments. The strata’s lateral variability highlights the depositional environments from a single time. And their vertical succession shows the changes in depositional environment through time. Nineteen volcanic layers are found in the Wilson
The 19 tephra of the Wilson Creek Formation are estimated to encompass ~66 to 15 ka (Benson et al., 1990, 1998; Zimmerman et al., 2006; Vazquez and Lidzbarski, 2012). They are numbered by their inverse emplacement: Ash 19 is the oldest; and Ash 1 is the youngest. Eighteen tephra are rhyolitic; one is basaltic. Seventeen of these rhyolitic tephra are thought to have erupted from the Mono Craters. Mammoth Mountain is the source of the other rhyolitic tephra, Ash 18. The basaltic tephra, Ash 2, reflects the sublacustrine eruption of Black Point (Lajoie, 1968).

Parts of the Wilson Creek Formation tephra are exposed as low as 1,955 m and as high as 2,145 m. Because unique characteristics define some tephra, their identification is unambiguous (Lajoie, 1968). Less distinct tephra need a succession of two or three tephra for unequivocal field identification.

During Wilson Creek time, Benson et al. (1990, 1998) suggested that Mono Lake exceeded a level of 2,035 m except for two intervals: between the deposition of Ashes 5 and 4; and between the deposition of Ashes 15 and 16. During the Ash 5-4 interval, these authors estimated that the lake fell from 2,075 m to 2,035 m. They, too, argued that the lake remained at this low level between 24-17 ka before rising to the lake's 2,155-m high stand ~15.5 ka. And they infer that the lake fell from this high stand to 1,965 m over a 4 kyr interval. One brief lake rise from 2,010 to 2,020 m at ~12.6 ka interrupted this lake fall. But the lake fell immediately after the 12.6-ka high stand to 1,965 m and fluctuated little during the next 8 kyr.
3.3 Methodology

The lake fluctuation record I present herein differs from previous studies. Three data underpin my record: (1) stratigraphic evidence from newly discovered as well as previously described sedimentary sequences; (2) geomorphic features that were formed by the lake’s transgressions and regressions; and (3) sixty U/Th and seven new $^{14}$C data from terrestrial material that help constrain the timing of the lake’s fluctuations. Nearly all the U/Th analyses were made on lacustrine carbonates of varying origins. The elevations of some algal carbonates and calcareous hardgrounds are near-direct lake level constraints (see Appendix). Algal tufa form in the shallow-water photic zone (Burne and Moore, 1987; Dupraz and Visscher, 2005; Dupraz et al., 2009; Franks and Stolz, 2009). Calcareous hardgrounds form where the groundwater table intersects the shoreline (Last and DeDecker, 1990; Harrison, 2017). I used the elevations of both deposits as lake level indicators. On the other hand, dates from calcite-cemented conglomerates and meteoric carbonates indirectly inform lake level. They provide minimum and maximum lake level constraints, respectively. Radiocarbon dates from charcoal and bones cemented in tufa provide minimum ages for lake transgressions. And dates on charcoal found in sediments serve as maximum estimates on the strata that enclose them. The carbonate U/Th data are displayed in Table 3.1, and the terrestrial macrofossil $^{14}$C data are displayed in Table 3.2.

3.4 Results

My data show that Mono Lake fell to its lowest elevation during the peak of the last glacial period. The elevational range of this epic low stand, termed the "Big Low", measures somewhere between 1,975 to 1,955 m. It is the most extreme low stand of the last glacial cycle. At the time of this
thesis, I do not have a direct date on this low stand. But the data I do have show that the Big Low occurred in the interval between 24.4 ± 0.2 ka and 20.5 ± 0.2 ka.

The rise of the lake to its 2,155-m high stand began by 20.5 ± 0.2 ka. The lake was at its highest level by 15.94 ± 0.05 ka. And it, then, fell to 2,010 m in two stages: the first—from 2,155 m to 2,075 m—occurred between 15.90 ± 0.05 ka and 15.07 ± 0.06 ka; the second—from 2,075 m to 2,010 m—occurred between 14.1 ± 0.1 ka and 13.8 ± 0.2 ka. Both falls of the lake corresponded with a 30% contraction of the lake's surface area. And both regressions were rapid.

No samples are found that date between 13.8 ka and 13 ka. Starting at approximately 13 ka, the lake rose from 2,010 m and reached 2,075 m by ~12.1 ± 0.1 ka. This lake rise terminated at 2,089 m. The precise timing of this high stand is unconstrained. The lake’s decline from this high stand is, however, well constrained. The lake fell as low as 1,982 m by 11.0 ± 0.5 ka. Additional data show that the lake fluctuated across 1,975-1,970 m during the time encompassing 11-9.5 ka.

### 3.5 Discussion

Mono Lake’s fluctuations from 20.5 to 10 ka parallel eminent records of tropical and mid-latitude hydroclimate (Wang et al., 2001; Dykoski et al., 2005; Deplazes et al., 2013; Figure 3.2). The tropical records show variations in the position of the Intertropical Convergence Zone (ITCZ). And the mid-latitude records indicate the strength of the east Asian monsoon. And they broadly coincide with the variations in North Atlantic climate—namely, those associated with abrupt warming and cooling (Heinrich, 1988; Broecker et al., 1992; Bond et al., 1992; Hemming, 2004; Grootes et al., 1993; Figure 3.2). For instance, periods of cooling in the North Atlantic that are associated with southerly excursions of ice-rafted detritus or sea ice—Heinrich Stadial 1, 17.5-14.7 ka, and Younger Dryas, 12.9-11.7 ka, for
example (Wang et al., 2001; Barker et al., 2009; Goni and Harrison, 2010; Rasmussen et al., 2014)—fit with Mono Lake transgressions, southward diversions of the intertropical convergence zone (Deplazes et al. 2013), and reductions in the strength of the east Asian monsoon (Dykoski et al., 2005; Wang et al., 2010). Anomalous warming in the North Atlantic—such as during the Bølling-Allerød, 14.7-12.9 ka, and the Holocene, 11.7 ka to present (Grootes et al., 1993; Björck et al., 1998; Wang et al., 2001; Rasmussen et al., 2014)—is contemporary with the converse: regressions of Mono Lake; northward migrations of the intertropical convergence zone; and intensifications of the east Asian monsoon. One departure from this correspondence is during the last ~1 kyr of Heinrich Stadial 1. All three records of hydroclimate pivot at 16 ka from their most extreme Heinrich Stadial 1 values. But I find no analogous shift in Greenland ice core records. The cause of this deviation is unclear. But the harmony between the hydroclimatic records suggests that there is a common conductor of the global hydrologic cycle.

The best-accepted hypothesis that explains western US hydroclimate during the last glacial cycle is commonly referred to as the “shifting westerlies”. This hypothesis, which follows conclusions from an earlier climate dynamics study (Antevs, 1952), advances the argument that the size of the North American ice sheet influences the position of the mid-latitude jet (e.g., Thompson et al., 1993; Bartlein et al., 1998). Because of the ice sheet’s permanence, these studies argue that the mid-latitude jet was perennially diverted to the south. And they suggest that this southward diversion of the mid-latitude jet promoted an increase in wetness across the Great Basin, causing its lakes to swell. Likewise, the model predicts that the mid-latitude jet retreated to the north as the ice-sheet waned during the deglaciation. This northward diversion of the jet was met with reduced wetness in the Great Basin, which forced lakes to contract. Thus lakes rose at the glacial maximum, and they fell during the deglaciation. These predictions, however, are not compatible with the pattern and timing of Mono Lake’s fluctuations: for the lake both rose and fell during the deglaciation and at the time of glacial maximum. Another reason to question the
shifting westerlies hypothesis is that it cannot explain the synchrony between Mono Lake’s fluctuations and shifts in other indices of mid-latitude and tropical wetness (Figure 3.2). Nor can the shifting westerlies hypothesis explain the Mono Lake’s correlation to North Atlantic climate. By these inconsistencies, I argue that the shifting westerlies is an invalid model to explain past hydroclimate in the Great Basin.

Well-established studies that link tropical and extratropical hydroclimate via the Hadley circulation afford an alternative hypothesis (Chiang and Bitz, 2005; Lee et al., 2011; Chiang and Friedman, 2012; Chiang et al., 2014). They show that northern hemisphere cooling, such as that observed during Heinrich stadials, deflects the ITCZ to the south. The response to this meridional shift of the ITCZ is a strengthening of the northern hemisphere’s winter Hadley cell, which amplifies the energy transport from the tropics to the subtropics (Lee et al., 2011; Chiang et al., 2014). This amplification is reflected, in part, by an intensification of the wintertime subtropical jet stream (Lee et al., 2011; Chiang et al., 2014). A stronger subtropical jet augments the meridional transport of tropical and subtropical moisture to east-central California (Lee et al., 2011; Chiang et al., 2014). Modelling experiments retrodict that a factor of three to four increase in precipitation will occur as a result of an intensification of the wintertime subtropical jet under climatic conditions comparable to Heinrich Stadial 1 (Lee et al., 2011; Chiang et al., 2014). They, too, show a concurrent weakening of the boreal summer monsoon. Thus the simulation of conditions that would give rise to a wetter Mono Basin, a weaker east Asian monsoon, and a southward deflection of the ITCZ during North Atlantic cooling episodes lends confidence to the Hadley circulation model (Figure 3.3). And its predictions for periods of abrupt North Atlantic warming—i.e., contractions of Mono Lake, strengthening of the east Asian monsoon, and a northward migration of the ITCZ—are consistent with the geologic evidence, too (Figure 3.3).

Comparisons of the Mono Lake record to other Great Basin lake records is challenging. No other lake fluctuation record from the Great Basin is as complete or well-resolved as the record I
present here. But where direct comparisons can be made, the Mono record is harmonious with lake records from the Searles and Lahontan Basins. Multiple salt deposits in the Searles Basin, which highlight intervals of drought, date to ~35-23 ka. There are at least two of these salt deposits that overlap with the timing of the Big Low. And the deglacial high stand of Searles Lake, which is dated to ~16.5 ka, also agrees with Mono’s high stand datum (Lin et al., 1998). Likewise, the high stand of Lake Lahontan—estimated to be 15.7 ka (Adams and Wesnousky, 1998)—corresponds with Mono Lake’s ~16 ka high stand. Furthermore, Lake Lahontan’s recession from its high stand, its subsequent rise during the Younger Dryas, and its fall in the early Holocene all correspond with the Mono record (Benson et al., 2013).

This harmony notwithstanding, a comparison between the Mono and Bonneville Lake records show a more complex relationship. Mono Lake and Lake Bonneville both transgressed and regressed 20-18 ka and 16-15 ka, respectively. But the two lakes were out of phase during other times—namely, Lake Bonneville appears to have rapidly rose when Mono Lake fell to the Big Low. Because Lake Bonneville bifurcated into two lakes with unique hydroclimatic signals after ~15 ka, a direct comparison is dubious (Oviatt, 2015). Together the evidence for varied hydroclimate signals across the Great basin implies distinct hydroclimatic regions during the last glaciation.

Global hydroclimate proxy records agree with the predictions made by the Hadley circulation model. This agreement suggests that the Hadley circulation model is the most tenable explanation for past hydroclimate shifts. And it asserts that the controller of Great Basin hydroclimate is the interhemispheric temperature gradient. In order to further test this hypothesis, greater chronologic precision and temporal resolution are needed. This necessitates more lake level studies.

The Big Low remains an enigmatic feature of the Mono Lake record. Too few data are available to precisely constrain its timing. But if the Big Low’s hydroclimatic forcings were consistent with those
of the deglacial and early Holocene, I suspect that this low stand coincided with the abrupt warming recorded in the Greenland ice-core record at ~23 ka (Rasmussen et al., 2014). Testing this hypothesis requires more field and analytical work.

3.7 References


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<th>Sample ID</th>
<th>Locality</th>
<th>Elevation (m)</th>
<th>$^{238}$U (ppb)</th>
<th>$^{232}$Th (ppt)</th>
<th>$^{234}$U (measured)</th>
<th>$^{232}$Th (corrected)</th>
<th>$^{234}$U (initial)</th>
<th>Th/U</th>
<th>Age (Ma)</th>
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<tr>
<td>ML1</td>
<td>eastern Sierra Rd</td>
<td>1890</td>
<td>5921 ± 15</td>
<td>24584 ± 496</td>
<td>0.1797 ± 0.0005</td>
<td>19077 ± 72</td>
<td>12157 ± 51</td>
<td>0.1762 ± 0.0006</td>
<td>1.97</td>
</tr>
<tr>
<td>ML2</td>
<td>Goat Ranch Rd</td>
<td>201109</td>
<td>5000 ± 10</td>
<td>1024 ± 39</td>
<td>0.1491 ± 0.0004</td>
<td>10475 ± 107</td>
<td>1024 ± 39</td>
<td>0.1479 ± 0.0005</td>
<td>1.96</td>
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Table 3.1: All carbonate U/Th data
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<tr>
<th>Sample ID</th>
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<th>232Th (ppt)</th>
<th>234U (ppb)</th>
<th>236U (ppb)</th>
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<th>Age (Ma)</th>
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Table 3.1: All carbonate U/Th data, continued
Sample ID
U (ppb)

230Th Age
(a)***
(corrected)

Locality

234
Uinitial
corrected

Age
(ka)corrected
b,c

Age
(ka)uncorrected

234
U
(measured)a

[230Th/238U]
(activity) b

232
Th
(ppt)

Elevation
(m)

11996 ± 44

12057 ± 44

232 ± 2

12060 ± 44

12387 ± 91

0.1285 ±
0.0004

210 ± 2

12079 ± 69

224 ± 1

12448 ± 91

225 ± 2

12935 ± 70

1490 ± 44

12473 ± 90

12139 ± 69

227 ± 2

12920 ± 104

10377 ±
18

12154 ± 69

12995 ± 70

221 ± 2

12663 ± 55

2030

13056 ± 56

225 ± 2

Upper Wilson
Creek

13118 ± 38

12891 ±
104
12724 ± 55

203 ± 2

12771 ± 44

14371 ± 88

15070 ± 60

205 ± 2

14489 ± 72

191 ± 2

14429 ± 88

203 ± 2

15103 ± 47

14530 ± 52

14547 ± 72

14192 ± 50

15187 ± 44

14568 ± 71

210 ± 2

15040 ± 60

12568 ±273

195 ± 2

14255 ± 50

191 ± 2

2030
2030
2030

29631 ±
600
21005 ±
422
9325 ± 191
8089 ± 165
9135 ± 185
14644 ±
171
3387 ± 176

14260 ± 50

14222 ± 52

14470 ± 107

211 ± 2
214 ± 2

14439 ± 100

14285 ± 52

216 ± 2

12200 ± 160

14290 ± 52

254 ± 3

14270 ± 230

14668 ± 49
14625 ± 51
12412 ± 122
14338 ± 228

208 ± 2

14533 ±
107
14502 ±
100
12344 ±
122
14335 ±
228

15073 ± 47

2493 ± 78

238

Table 3.1: All carbonate U/Th data, continued

ML11UWC-7A2
ML11UWC-7B
UWC2-1A
UWC2-1B
2030

9438.5 ±
14.6
7966 ± 14

2030

2070

8493 ± 19

201.5 ±
1.7
203.1 ±
1.9
205.0 ±
1.5
207.4 ±
2.1
245 ± 3
200 ± 2

15133 ± 47

0.1304 ±
0.0009
0.1288 ±
0.0007
0.1380 ±
0.0005
0.1379 ±
0.0003
0.1349 ±
0.0004
0.1543 ±
0.0004

2070

2070

1291 ± 44
1165 ± 42

183.3 ±
1.5
197 ± 2

217.7 ±
1.4
218.9 ±
1.5
212.8 ±
1.5
216.7 ±
1.6
183.1 ±
1.5

Upper Wilson
Creek
Upper Wilson
Creek
Upper Wilson
Creek
Upper Wilson
Creek
Upper Wilson
Creek
Hwy 167

2070

2075

13057 ±
27
13034 ±
31
8211 ± 14

20343 ±
415
185-1 ± 380
15846 ±
341
3240 ± 244

0.1541 ±
0.0004
0.1498 ±
0.0004
0.1500 ±
0.0007
0.1478 ±
0.0004
0.1483 ±
0.0004
0.1522 ±
0.0004
0.1550 ±
0.0004
0.1344 ±
0.0011
0.1483 ±
0.0022

11918 ±
47
4042.1 ±
10.7
11660.3 ±
28.8
8304.8 ±
11.9
10728.9 ±
19.5
8555.5 ±
13.2

Hwy 167
Hwy 167

2075

8204 ± 18

UWC21C1
UWC21C2
ML10MC1-1A
ML10-6,
ML10MC1-1B
2A
Hwy 167

2075

5421 ± 24

2075

2075

20703 ±
210

Goat Ranch
Rd cutoff Rd
Goat Ranch
Rd cutoff Rd
Goat Ranch
Rd cutoff Rd
Goat Ranch
Rd cutoff Rd
Goat Ranch
Rd cutoff Rd
Goat Ranch
Rd cutoff Rd

2075

Mono
102A
GR13-2-1
(EOS)
GR13-2-1
(Hai)
GR13-2-2
(EOS)
GR13-2-2
(Hai)
ML201207
-GR3-07-1
ML201207
-GR3-07-2

84


<table>
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<tr>
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<th>Locality</th>
<th>Elevation (m)</th>
<th>238U (ppb)</th>
<th>232Th (ppt)</th>
<th>234U (measured)</th>
<th>[230Th/238U] (activity)</th>
<th>Age (ka)</th>
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<tr>
<td>ML201207-GR3-07-3</td>
<td>Goat Ranch Rd cutoff</td>
<td>2075</td>
<td>12936 ± 48</td>
<td>670 ± 79</td>
<td>208 ± 2</td>
<td>0.1490 ± 0.0009</td>
<td>1249 ± 0.09</td>
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<td>ML201207-GR3-07-4</td>
<td>Goat Ranch Rd cutoff</td>
<td>2075</td>
<td>12937 ± 48</td>
<td>2070 ± 100</td>
<td>226 ± 2</td>
<td>0.1501 ± 0.0010</td>
<td>14191 ± 105</td>
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<tr>
<td>ML201207-GR3-07-5</td>
<td>Goat Ranch Rd cutoff</td>
<td>2075</td>
<td>12938 ± 48</td>
<td>11708 ± 278</td>
<td>210 ± 2</td>
<td>0.1515 ± 0.0014</td>
<td>14526 ± 143</td>
</tr>
<tr>
<td>ML201207-GR3-07-6</td>
<td>Goat Ranch Rd cutoff</td>
<td>2075</td>
<td>12939 ± 48</td>
<td>3697 ± 113</td>
<td>209 ± 2</td>
<td>0.1496 ± 0.0009</td>
<td>14345 ± 99</td>
</tr>
<tr>
<td>ML201207-GR3-07-7</td>
<td>Goat Ranch Rd cutoff</td>
<td>2075</td>
<td>10411 ± 56</td>
<td>2141 ± 125</td>
<td>209 ± 2</td>
<td>0.1524 ± 0.0013</td>
<td>14642 ± 139</td>
</tr>
<tr>
<td>ML201207-GR3-07-8</td>
<td>Goat Ranch Rd cutoff</td>
<td>2075</td>
<td>9824 ± 36</td>
<td>2441 ± 89</td>
<td>207 ± 2</td>
<td>0.1508 ± 0.0009</td>
<td>14497 ± 92</td>
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<tr>
<td>ML201207-GR3-07-9</td>
<td>Goat Ranch Rd cutoff</td>
<td>2075</td>
<td>9500 ± 48</td>
<td>5261 ± 153</td>
<td>208 ± 3</td>
<td>0.1520 ± 0.0012</td>
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<td>SN-3-2</td>
<td>eastern Sierra Nevada</td>
<td>2145</td>
<td>6002 ± 12</td>
<td>732 ± 23</td>
<td>224.2 ± 2.0</td>
<td>0.1674 ± 0.0005</td>
<td>15966 ± 55</td>
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<td>UWC5-1A</td>
<td>Upper Wilson Creek</td>
<td>2030</td>
<td>9520 ± 15</td>
<td>5518 ± 1107</td>
<td>173 ± 1</td>
<td>0.2051 ± 0.0005</td>
<td>20847 ± 59</td>
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<tr>
<td>UWC5-1B</td>
<td>Upper Wilson Creek</td>
<td>2030</td>
<td>35340 ± 715</td>
<td>772 ± 16</td>
<td>174 ± 1</td>
<td>0.2022 ± 0.0008</td>
<td>20509 ± 90</td>
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<tr>
<td>S49</td>
<td>Big Low type-locality</td>
<td>1982</td>
<td>1234 ± 5</td>
<td>210333 ± 4312</td>
<td>160.5 ± 3.0</td>
<td>0.2336 ± 0.0014</td>
<td>24379 ± 182</td>
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Table 3.1: All carbonate U/Th data, continued
Table 3.4: All carbonate/feldspar data, continued.

![Image of table with data](image-url)
<table>
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<th>Sample Name</th>
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<th>Location</th>
<th>Lat./Long.</th>
<th>Elevation (m)</th>
<th>Material</th>
<th>C-14 Age (yrs BP)</th>
<th>14C Age (yrs BP)</th>
<th>IntCal13-calibrated age</th>
<th># replicate</th>
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<td>2013-10-38-119-87</td>
<td>170035</td>
<td>38.057548, -119.141122</td>
<td>2,076</td>
<td>bird bone</td>
<td>10210 ± 35</td>
<td>11920 ± 90</td>
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<td>2013-09-86-06-05</td>
<td>163510</td>
<td>38.020545, -119.132296</td>
<td>1,968.65 charcoal</td>
<td>8685 ± 40</td>
<td>9630 ± 60</td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>2013-09-86-06-06</td>
<td>163512</td>
<td>38.020545, -119.132296</td>
<td>1,968.65 charcoal</td>
<td>8495 ± 30</td>
<td>9510 ± 20</td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>
Figure 3.1. NASA satellite image showing the Mono Basin. Blue-colored lines show Mono Lake’s perennial tributary streams: (1) Rush; (2) Lee Vining; (3) Post Office; (4) Mill; and (5) Wilson. Marker (6) is the basaltic cinder cone Black Point. The orange-colored areas highlight the Mono and Inyo Craters.
Figure 3.2. A compilation of four data sets showing disparate archives of hydroclimate 25-8 ka. Series “a” shows this study’s Mono Lake fluctuation record. Series “b” shows the oxygen
isotope record of the East Asian monsoon from two caves: Hulu (orange, Wang et al., 2001); and Dongge (black, Dykoski et al., 2005). More negative oxygen isotope values indicate a stronger monsoon. More positive values are inferred to reflect a weaker monsoon. Series “c” shows the record of the reflectance of sediments in the Cariaco Basin, which lies near the Venezuelan coast (Deplazes et al., 2013). The reflectance indicates the relative proportion of sediments deposited via terrestrial runoff and biogenic strata. It is used to infer the mean position of the ITCZ. A greater reflectance indicates less runoff and a more southerly position of the ITCZ; the converse is reasoned for a decrease in reflectance. Series “d” shows an oxygen isotope record taken from the Greenland Ice Sheet (GISP2, Grootes et al., 1993). It is inferred to reflect climatic conditions in the North Atlantic. Periods of cooling are recorded as more negative oxygen isotope values; warming is indicated by less negative oxygen isotope values. The four records are harmonious. When Mono Lake is rising—and by extension when east-central California is wetter—the East Asian monsoon is weakening, the ITCZ has shifted to the south, and the North Atlantic is defined by inordinate cooling. On the other hand, it appears that Mono Lake is falling and the region is drying when (1) the East Asian monsoon is strengthening, (2) the ITCZ diverts to the north, and (3) the North Atlantic warms. A pivot exists in the Cariaco Basin, Hulu cave, and Mono Lake records at 16 ka (highlighted by the arrow) that is not apparent in the Greenland ice core.

Figure 3.3. Cartoon of four harmonious, yet disparate, hydroclimate records. Their harmony is described here after Chiang et al., (2014). Large lakes in the Great Basin are fed by a strengthened wintertime subtropical jet. This occurs as a response to a southward diversion in the ITCZ. This teleconnection is mediated by the Hadley circulation. The meridional shift of the ITCZ is a reflex to surface pressure anomalies in the North Atlantic owing to an anomalous extent of winter sea ice. The common indicator of change amongst these elements is the interhemispheric temperature gradient. And when the gradient is in favor of Northern Hemisphere cooling, the Great Basin gets wetter. And when the Northern Hemisphere warms relative to the Southern Hemisphere, it gets drier in the Great Basin.
Chapter 4: The Big Low: an extreme low stand of Mono Lake during the Last Glacial Maximum

4.1 Abstract

The coolest chapter of the last glacial period, termed the Last Glacial Maximum (LGM), encompassed ~27-19 ka. During this time, it is thought that elevated wetness sustained large lakes across the Great Basin of the western US, in spite of the sparse data and discontinuous time series supporting the interpretation. Here I show evidence for an extreme low lake level in the Mono Basin, a hydrologically closed basin in east-central California, that occurred during the LGM and sometime in the interval ~24.4-20.5 ka. The timing of this low stand corresponds with summer temperature minima, suggesting that the fall was due not to an increase in evaporation but due to a decrease in precipitation. Both the documentation of a low stand during glacial maximum conditions and the inference that precipitation must have been reduced are contrary to previous published interpretations from model and paleoclimatic data. These discrepancies raise significant questions about our understanding of the regional expression and forcing of hydroclimate across the western US during the LGM. Because of this period’s importance to ground-truthing climatic models, additional evidence on the geographic extent of this unexpected result is essential.
4.2 Introduction

Continental ice sheets and mountain glaciers reached their largest extents during the so-called Last Glacial Maximum (LGM, 27-19 ka, Clark et al., 2009), an interval of insolation and temperature minima (Berger and Loutre, 1991; Dahl-Jensen et al., 1998). A nearly 125-year-old problem in paleoclimate studies centers on how lakes in the Great Basin of the western US responded to the extreme climatic changes during this time (Figure 4.1; Russell, 1885, 1889; Gilbert, 1890; Putnam, 1950; Antevs, 1952; Antevs, 1955; Broecker and Orr, 1958; Morrison, 1964; Morrison and Frye, 1965; Smith, 1968; Benson and Thompson, 1987; Benson et al., 1990; Allen and Anderson, 1993; Benson et al., 1997; Oviatt, 1997; Benson et al., 1998; Lin et al., 1998; Benson et al. 2003; Bacon et al., 2006; McGee et al., 2012; Ibarra et al., 2014; Reheis et al., 2017). Fragmented lake level evidence, disparate proxy data, and difficulties with dating preclude definitive conclusions. As a result, dissimilar and contradictory mechanisms have been proposed to explain lake fluctuations (Antevs, 1952; Kutzbach and Wright, 1985; Benson et al., 1990; Bartlein et al., 1998; Benson et al., 1998; Benson et al., 2003; Lyle et al., 2012; Munroe and Labs, 2012; Benson et al., 2013; Ibarra et al., 2014; Oster et al., 2015; Lora et al., 2017). Ultimately this quandary makes it impossible to evaluate whether the current suite of general circulation models can accurately simulate past western US hydroclimate.

Mono Lake lies in an actively-extending basin in east-central California (Figures 4.1-4.2; Pakiser, 1960; Bursik and Sieh, 1989). Its current surface elevation is ~1,945 m. The high-relief Sierra Nevada bounds its western edge, and its other three margins are marked by volcanic ranges of low to moderate relief---termed the Bodie Hills, Cowtrack Mountains, and Long Valley Caldera (Figure 4.2). The southern margin of the Mono Basin is punctuated by the Mono-
Inyo Craters, a chain of thirty or more volcanic domes and coulees, the youngest of which is ~600-years-old (Figure 4.2; Sieh and Bursik, 1986; Bailey, 1989; Hildreth, 2004).

Mono Lake’s principal inflow is delivered by its Sierran tributary streams (Voster, 1985). These tributary streams are, from north to south, Mill, Post Office, Lee Vining, Parker, and Rush. Their runoff is sustained by spring and summer melting of Sierran snowpack, which is delivered by cool-season westerly storms (Serreze et al., 1999).

Because of the basin’s hydrologic closure, the surface area of Mono Lake reflects the balance between precipitation and evaporation. Precipitation constitutes inputs from inflow and direct precipitation. When precipitation exceeds evaporation, lake surface area expands. But when evaporation is greater than precipitation, the lake’s surface area shrinks. The basin’s hypsometry shows a near-linear correlation between lake surface area and lake elevation (Figure 4.3). Thus lake level rises and falls are a result of an increase or decrease in precipitation minus evaporation.

The Mono Basin's Late Pleistocene lithostratigraphic unit is the Wilson Creek Formation (Lajoie, 1968). It is exposed along stream-, wave-, and road-cuts (Lajoie, 1968). The most extensive exposures are along the basin’s deeply incised streams--namely, Mill, Lee Vining, and Rush Creeks. The sedimentary sections observed along their canyon walls show interfingering and cross-cutting glacial, littoral, fluvial, and lacustrine sediments (Lajoie, 1968). Together, these sedimentary sequences provide the necessary data to constrain a detailed record of lake fluctuation (Lajoie, 1968).

Tephra deposits found in the Wilson Creek Formation afford a relative time series of lake level: for they are definite time markers. In total, 19 tephra were documented in lacustrine silts in the original description of the Wilson Creek Formation (Figure 4.4; Lajoie, 1968). They are
numbered by their inverse emplacement: 19 is the oldest; and 1 is the youngest. One tephra, Ash 2, is basaltic, and it is derived from Black Point (Lajoie, 1968). The remaining 18 tephra are rhyolitic. Seventeen of these are from Mono Craters (Marcaida et al., 2014). The exception is Ash 18, which was erupted from Mammoth Mountain (Marcaida et al., 2014).

During Wilson Creek time, three periods of high lake level and two periods of low lake level are inferred from geochemical data (Zimmerman et al., 2011). The intervals of high lake level encompass the deposition of Ashes 19-16, Ashes 15-8, and Ashes 7-1. And the intervals of low lake level include time time of Ashes 16-15 and Ashes 8-7. My interpretations are in accord with the interpretations presented in the Zimmerman et al., (2011) study; however, in this paper, I present new data on an exceptional low stand of the lake during a period of relatively high lake level.

Here I document the existence of extreme low stand(s) of Mono Lake in the interval between ~24 and ~21 ka. My interpretations are underpinned by two components: (1) a careful reassessment of stratigraphy at several key localities that comprise inter-fingering fluvial, glacio-fluvial, littoral, and lacustrine sediments and their contained tephras; and (2) U/Th dates on lacustrine and meteoric carbonates.

The motivation for the study presented herein was the discovery of an enigmatic deposit along the eastern canyon walls of Mill Creek (Figures 4.2, 4.5, and 4.6). This deposit, which I describe in the following section, is the fill of a cut-and-fill deposit. It comprises a sequence of sand and gravel that includes poorly sorted cobble gravel, and it overlies and underlies Wilson Creek Formation strata. The gravel is fluvial. And its basal elevation is 1,982 meters. This fluvial gravel deposit caught my attention because it is nearly 30-vertical meters lower than what previous authors conjectured to be Mono Lake’s greatest low stand of glacial times (Lajoie,
1968; Benson et al., 1990). I informally refer to this fluvial deposit at Mill Creek as the “Big Low type-locality”.

4.3 The Big Low type-locality

Two nested cut-and-fill sequences that truncate thinly-laminated silts occur along the eastern channel walls of Mill Creek (Figures 4.2, 4.5, and 4.6). There are no published descriptions of this outcrop. I describe it from observations I made across four-vertical meters and 35-lateral meters of exposure.

The first cut-and-fill sequence (G1) is divided into four units (1-4) (Figure 4.12): a lower composite unit comprising well sorted, planar and wavy sands (G1-1) and massive, poorly sorted gravel that ranges up to cobbles in size (G1-2) (Figure 4.11); a medial unit comprising very well sorted, cross-laminated, pebble conglomerate (G1-3) (Figure 4.13); and an upper unit of moderately sorted sands and granules and poorly-sorted gravel up to cobbles in size (G1-4) (Figure 4.12). The lower unit is underlain by thinly-laminated silts and clayey-silts of the Wilson Creek Formation (S1) and Ashes 15-8 (Figure 4.8). The contact below the sand and gravel sequence is erosional. Intraformational clasts of Ash 8 and an intraformational block of highly-deformed lacustrine silt (S2) containing Ashes 7-5 are found in the lower fill unit (Figures 4.14-4.15). I sampled calcite crystals that cemented the conglomerate of the medial unit (G1-3) for U/Th analysis.

The second cut-and-fill sequence truncates the first. Its basal contact is curvilinear (Figure 4.9). Like the first sequence, it incises into the thinly-laminated silts that contain Ashes 15-8, too (Figures 4.8-4.9). The strata that compose the second cut-and-fill sequence are
threelfold. The lowest exposure is comprised of poorly sorted gravel up to cobble in size (G2, Figure 4.8). The G2 gravel abuts similar strata that composes the first fill sequence (Figure 4.8). Overlying the G2 gravel is a two-part sequence that abuts and overlies the first cut-and-fill deposit (Figure 4.9). This bipartite sequence comprises thinly- to thickly-laminated and variably deformed silt and very well sorted sands, which are in turn overlain by laminated silts containing Ashes 4-1 (S3, Figure 4.10). Poorly sorted gravel up to cobble in size overlies Ash 1.

These stratigraphic data constitute the lowest elevation of recognized evidence for riverine strata in the Wilson Creek Formation. The observations described above require that the lake regressed to below 1,982 m following the deposition of Ash 5. During the time encompassing this “Big Low” low stand, a stream incised through Wilson Creek lacustrine strata. This incision must have formed a canyon. I interpret the curvilinear contact between the two cut-and-fill sequences to be the surface of this ancient canyon. The lacustrine silts that overlie this paleosurface--and the tephra they contain--require that the lake transgressed from the Big Low to some elevation above this deposit prior to the deposition of Ash 4. I, therefore, deduce that the Big Low occurred between the deposition of Ash 5 and Ash 4.

The cementation of the littoral embankment with calcite is enigmatic because I do not find it in any other deposit. Because the calcite cement is exclusive to the littoral embankment, and because the strata are not different than those below or above it, I infer that the calcite likely formed as the littoral embankment was being deposited. By this reasoning, the U/Th datum on the calcite would closely constrain the timing of the littoral embankment’s deposition. And because the second cut-and-fill sequence cross-cuts the embankment, the U/Th datum is a maximum age for when the lake transgressed from the Big Low.
4.4 Observations from upstream of the Big Low type-locality

Approximately 200 m upstream of the Big Low locality, and also along the eastern channel walls of Mill Creek, the full set of Wilson Creek Formation tephra are found in thinly- to thickly-laminated silts that are underlain by poorly sorted, pre-Wilson Creek gravel (Figure 4.5). The elevation of the Wilson Creek Formation here is 1,986 to 1,993 m. Within the interval between Ash 5 and 4, I found a deposit that I infer to represent the expression of the Big Low (Figure 4.16). This deposit comprises very well sorted sands that are underlain by a horizon of pebbles and cobbles. The basal contact of this deposit, which I measured to be ~1,991 m, is sharp. The contact cross-cuts a convoluted interval of silt that contains Ashes 7-5.

I interpret the poorly sorted cobble gravel as fluvial gravel, the thinly-laminated silts as lacustrine silts, and the very well sorted sands as littoral sands and the cobble and pebble clasts as lag deposits. I, therefore, conclude that the contact between the littoral sands and the lacustrine silt it overlies is a wave-cut disconformity. I interpret that the lake fell below the 1,991-m disconformity within the time encompassing the deposition of Ashes 5-4.

4.5 Wilson Creek type-locality

Because the Big Low type-locality requires the lake to have fallen below 1,982 m, there must be a correlative disconformity between Ashes 5 and 4 in the Wilson Creek type locality: for the sediments exposed along lower Wilson Creek are as low as 1,970 m and as high as 1,999 m. Previously published observations document the presence of deformed silts and sand lenses between Ashes 5 and 4 (Figures 4.2 and 4.17; Lajoie, 1968; Benson et al., 1998).
The sands found between Ashes 5 and 4 are very well sorted (Figure 4.18). Their lowest exposure is 1,975 m; their highest exposure is 1,989 m. They are underlain by planar parallel, thinly- to thickly-laminated silts that are intercalated with Ashes 19-5. And they are immediately overlain by a one-meter-thick interval of up-thrusted, thinly-laminated silts (Figures 4.19-4.20). The deformed laminae dip moderately to the north. These deformed silts are overlain by planar parallel, thinly-laminated silts intercalated with Ashes 4-1 (Figures 4.19-4.20).

Lajoie considered the sands to reflect deposition by sublacustrine mass wasting. But a different study suspected that they represented a disconformity between Ashes 5 and 4 (Benson et al., 1998). Though it is conceivable that these sands are sublacustrine mass wasting deposits, the most parsimonious explanation that is consistent with the other stratigraphic data along Mill Creek is that the sands are littoral sands that were deposited on a surface of wave planation. The low-angle, lakeward dip of this wave-cut disconformity is further consistent with this interpretation. If correct, this interpretation would extend the elevational limit of the Big Low from 1,982 m—which I concluded from the Big Low type-section—to 1,975 m.

4.6 Upper Wilson Creek

Where Cemetery Road crosses Wilson Creek (Figures 4.2 and 4.17) and approximately two kilometers upstream of our observed sections at the Wilson Creek Formation type-locality, Lajoie (1968) reported on an eight-meter-thick sequence (labeled as II-E) he measured that comprises well sorted sands and gravel that are overlain by silts containing Ashes 4-1. Lajoie, too, found physico-chemical tufa mounds rooted in the sands and gravel. Here I report on the same eight-meter-thick sequence with new observations.
My observations of Lajoie’s II-E show a sedimentary sequence comprising three successive units. The lower unit comprises tabular, cross-laminated, moderately-well to well sorted, granule gravel and sands that contain abundant rounded pumice lapilli. It is overlain by the middle unit, which comprises six meters of unconsolidated, very well sorted sands that interfinger with poorly sorted pebble conglomerate and poorly sorted pebble gravel. A meter-and-half of thickly- to thinly-laminated silts intercalated with Ashes 4-1 compose the upper unit.

I found spring tufa mounds rooted in the basal contact of the third unit. I sampled two physico-chemical calcite crystals from one of these tufa mounds for U/Th analysis (Figure 4.21). This tufa mound is overlain by Ashes 4-2 (Figures 4.22-4.23). Ash 1 is missing from the sedimentary sequence overlying this tufa mound; however, it is clear that Ash 1 was eroded away because I observed it in adjacent sedimentary sequences. I dated the calcite I sampled using U/Th analysis.

I interpret the tabular cross-bedded sand and granule gravel as a littoral embankment, the well sorted sands from the second unit as littoral sands, the pebble gravel and conglomerate that interfinger with the littoral sands as fluvial strata, and the laminated silts as hemipelagic lacustrine silts.

I am unable to identify the unknown tephra in the first unit at this time, but I reason that the abundance of pumice lapilli are characteristics consistent with Ash 7 or Ash 11—the only two Wilson Creek tephra found in the northwestern Mono Basin that contain pumice lapilli (Lajoie, 1968).

I deduce that this tripartite sedimentary sequence reflects a period of lake regression followed by a period of lake transgression. I interpret the littoral embankment to be deposited at lake level. And I interpret the overlying interfingerling littoral and fluvial strata to indicate that
the lake was oscillating at their approximate elevation. And because the deposition of fluvial strata is restricted to embayed delta trenches during lake transgressions (Stine, 1987), and because the sediments that compose this sequence were not deposited in an incised canyon, I reason that this sedimentary sequence reflects a period of lake regression and not lake transgression. I cannot determine the specific time of this lake regression; however, I do know that it must be prior to the deposition of Ash 4. I infer that the change from littoral and fluvial sediments to lacustrine silts indicates that the lake rose after they were deposited. And because Ash 4 overlies the change in strata from lake regressive to lake transgressive deposits, I reason that the lake rise initiated prior to the deposition of Ash 4. And based on the evidence presented thus far on the Big Low—that the lake rose from at least 1,975 m during the interval between Ashes 5 and 4—I conclude that there is a wave-cut disconformity present in this sedimentary sequence. I argue that because the strata below the silts were deposited during a time of lake regression, the most plausible horizon to mark the disconformity would be the contact between lacustrine silts and the littoral sands and gravel. And because the tufa mounds are rooted at this contact, their formation, which I infer was a result of groundwater degassing or lake- and ground-water mixing, is likely contemporaneous with the lake’s rise.

Although I cannot directly correlate each depositional unit in the context of the Big Low, the stratigraphy is consistent with my other findings—namely, that the lake fluctuated to a low stand following the deposition of Ash 5 and rose again prior to the deposition of Ash 4. Thus the time when the lake rose from the Big Low to 2,030 m can be constrained by a U/Th date on the tufa mound that was overlain by Ash 4.
4.7 Lower Bridgeport Creek

The lowest elevation outcrop of the Wilson Creek Formation that I have recognized measures from 1,955 to 1,960 m. It is located approximately one km from the present shore of Mono Lake (~1945 m) along the incised walls of the lowermost reaches of Bridgeport Creek (Figure 4.2). The sedimentary sequences there that contain Ashes 15 through 2 are cut and filled with Holocene deposits (Scott Stine, personal communication). At the specific sedimentary sequence described here, I find a continuous sequence of planar parallel, thinly- to thickly-laminated silt intercalated with Ashes 8-3 (Figure 4.24). Between Ashes 5 and 4, many of the laminae are marked by calcareous horizons. I found no sand or gravel deposits and no obvious disconformity in this interval.

The simplest explanation of these observations is that this exposure was continuously flooded by the lake throughout the interval encompassing the deposition of Ashes 15-2. I infer, therefore, that the lake did not fall as low as 1,955 m during the Big Low.

4.8 Israel Russell’s Tufa Crags

A 17-meter-high tufa deposit, termed the Tufa Crags (Russell, 1989), is exposed along a west-facing, wave-cut cliff (Figure 4.25). Thinly-laminated silts containing Ashes 15-1 abut the Tufa Crags. I also find silts within the gaps and interstices of the tufa. Within one of these interstices, I found two calcite deposits that were akin to those found in caves (Figure 4.26). One deposit grew from the roof of the interstice, and its structure appeared to droop downwards from the roof. The second calcite, which was on the floor of the interstice, grew upwards like a
stalagmite. Both deposits cut across Ash 5. I sampled the calcite deposit that formed as a stalagmite for U/Th analysis. Its elevation is 1,990 m.

I interpret the tufa mounds as sublacustrine spring deposits that formed prior to the deposition of the oldest tephra it is overlain by. I do not argue that the tufa was formed in one generation, and in fact, I reason that there were likely multiple episodes of tufa formation by their varying textures. Further study in this area could be fruitful for understanding ancient fluctuations of Mono Lake.

The carbonate deposits’ relationship to Ash 5 requires two pairs of observations and interpretations: that the calcite cross-cuts--and is therefore younger than--Ash 5; and that based on the macroscopic texture of the calcite, it formed in a subaerial, cave-like environment. By these interpretations, the lake was below 1,990 m when the calcite was precipitated. These findings are consistent with our elevational interpretations on the Big Low. Thus the U/Th age on the calcite that cross-cuts Ash 5 provides two constraints: a minimum age for Ash 5; and a maximum age for the Big Low.

### 4.9 Cottonwood Canyon

A four-meter-thick sedimentary sequence exposed along the eastern canyon walls of Cottonwood Creek (Figure 4.2) includes Ashes 8-5 in silts and clays. I measured the base of Ash 8 to be ~2,007 m. I found clays, coarse silts, and two beds of medium sands between Ashes 8 and 7. The lower bed is nine-cm-thick. It has granules at its base, and it fines upwards. The upper bed is two-cm-thick, and it has rounded pumice lapilli. Ashes 7-5 are intercalated in planar parallel, thickly- to thinly-laminated fine to coarse silts.
I use the Cottonwood Canyon sequence to place a lower limit on the elevational range of the lake during the time when Ash 5 was deposited. Based on the Cottonwood Canyon data and the aforementioned sedimentary sequences on the Big Low, I estimate the minimum elevational fluctuation of Mono Lake during the fall to the Big Low and the rise thereafter. I interpret the silts and clays as hemipelagic lacustrine sediments, both sandy beds as littoral deposits, and the granule gravel at the base of the lower sands to be a lag deposit. Thus I infer that the granules and sands in the lower bed are littoral sediments deposited onto a wave-cut disconformity. This requires that the lake fell to below ~2,007 m during the time interval encompassing Ashes 8 and 7. I do not, however, find evidence of a similar disconformity between Ashes 7-5. And I reason, therefore, that the lake did not fall below ~2,007 m during that time interval, Thus the data suggest that lake level was at least 2,007 m at the time Ash 5 was deposited.

4.10 On the lower elevational limit of Ash 4

In order to determine the elevational limit that the lake rose to after the Big Low and prior to the deposition of Ash 4, I measured the highest elevation of Ash 4 in lacustrine silts. I found Ash 4 in lacustrine silts along the canyon walls of Rush Creek and Bridgeport Creek at 2,052 m. Additional confidence in this elevational estimate is supported because these two localities are from opposing sides of the basin. Thus the lake rose from the Big Low--with an elevational range between 1,975 m and 1,957 m--to 2,052 m at the time Ash 4 was deposited.
4.11 Carbonate U/Th geochronology (see Table 4.1)

I present U/Th data from two calcite samples from the Upper Wilson Creek site (UWC5-1A and -1B, Figure 4.21), two from the Tufa Crags site (S549 and S550, Figure 4.26), and one from the Big Low site (GA-S49, Figure 4.13). I selected calcite that was dense and white, and apparently free of detritus. I sonicated the samples in ultra-pure water. And I used a carbide dental tool to mill ~1 mg of powder from the samples. I separated and purified the uranium and thorium in the sample following a methodology similar to Edwards et al., (1987) and Cheng et al., (2000). I measured the uranium and thorium isotopes by inductively-coupled mass spectrometry using a Thermo-Scientific Neptune 2 at Earth Observatory of Singapore. I assume that the present lake water $^{230}$Th/$^{232}$Th, which is $10 \pm 5 \times 10^{-6}$ (Anderson et al., 1982), is a reasonable estimate of the correction for calcite deposits of Wilson Creek age. The results of my analytical experiment are listed in Table 4.1. Included in this table are analyses made prior to this study by Dr. Xianfeng Wang in Dr. Larry Edward’s lab at the University of Minnesota.

The U/Th analyses yielded dates for UWC5-1A and -1B of $20,207 \pm 193$ a and $20,463 \pm 237$ a, respectively. This implies that Ash 4 is younger than 20.5 ka. And it suggests that the lake rose up to 2,030 m from the Big Low by this time, too.

The U/Th experiments on S549 and S550 from the Tufa Crags yielded ages of $24,478 \pm 329$ a and $25,106 \pm 537$ a, respectively. Because both samples cross-cut Ash 5, I conclude that Ash 5 must be older than 25.1 ka. And because the calcite is meteoric in origin, I deduce that the lake must have been lower than the elevation of the calcite samples--1,990 m--as early as 25.1 ka and as late as 24.5 ka. And I conclude that the chronologic constraints imply that the lake fell to within 15 to 34 m of the Big Low by 25.1 ka.
Using the modern lake composition to correct for initial thorium, my U/Th age on the calcite sample that cross-cuts the littoral embankment is 22,219 ± 1493 a. Its uncorrected U/Th age, 24,379 ± 182 a, is a maximum age. The deposition of the calcite as well as the littoral embankment, therefore, can be no older than ~24.4 ± 0.2 ka. And the Big Low is thus younger than ~24.4 ka ± 0.2 ka. This estimate is consistent with the inference that I made with the ~25 ka and ~24.5 ka calcite from the Tufa Crags.

4.12 Discussion

4.12.1 Local implications of the Big Low. Lajoie estimated that the lake regressed to 2,011 m during the time encompassing the deposition of Ash 5 and Ash 4. My estimation for the elevational range of the Big Low is that it was at least as low as 1,975 m but probably not as low 1,955 m (see summary in Figure 4.27). The new estimate presented here is at least 45 m lower than Lajoie’s estimate as well as the interpretations presented in the most recent published rendition of the lake’s Late Pleistocene fluctuation (Benson et al., 1990).

The Big Low’s elevational range brings attention to two matters. First, the Big Low represents the lowest elevations the lake fell to during Wilson Creek time. Thus the driest time during the last 50 kyr of the glaciation was at the time of glacial maximum. This is contrary to conventional wisdom, which suggests that the Last Glacial Maximum was a period of elevated wetness (Antevs, 1952; Kutzbach and Wright, 1985; Bartlein et al., 1998). The second point that merits consideration is that the Big Low is similar to elevations occupied by Mono Lake during the late Holocene--1,981 to 1,941 m, (Stine, 1990). Thus not only does the Big Low represent a
period of extreme dryness during Wilson Creek time, it also demonstrates that the glacial hydroclimate was, for a duration of time, similar to the wettest times recorded during the relatively warm Holocene interglacial. More significantly, the chronologic limits of the Big Low, 24.3 – 20.5 ka, suggest that the low stand occurred during the coolest interval of the last glacial cycle, the Last Glacial Maximum. The temperature minimum of the Last Glacial Maximum, I reason, drove evaporation rates to fall to a minimum, too. And if evaporation rates were reduced during the time of the Big Low, only a reduction in precipitation could cause a negative water balance. Thus I argue that the lake level fall during the Big Low was forced by a reduction in precipitation.

4.12.2 Big Low events in other Great Basin lake systems? Without evidence from other lake basins to corroborate the Big Low, it is a reasonable hypothesis to suggest that the Big Low reflects local rather than regional hydroclimatic forcing. If this hypothesis is true, the Big Low is an anomaly to be disregarded when placing it in the context of Great Basin hydroclimate; however, in a detailed reading of past—and often ignored—lake fluctuation literature, I’ve found corroborating evidence from two other Great Basin lakes (Figure 4.27): in the Pyramid Lake sub-basin of the Lake Lahontan system; and in the Searles Basin.

Searles Lake is a hydrologically closed basin that is located in southeastern California (Gale, 1914; Figure 4.1). Sedimentary cores from the center of the Searles Basin show seven salt layers (termed S1 to S7, from oldest to youngest) and six mud layers (termed M2 to M7, from oldest to youngest) during the time encompassing ~35-23 ka (Flint and Gale, 1958; Rubin and Alexander, 1958; Smith, 1962; Smith and Haines, 1964; Stuiver, 1964; Smith, 1979; Lin et al.,
1998). These strata necessitate repeated episodes of wetting and drying of the Searles Basin during the Last Glacial Maximum (Flint and Gale, 1958; Smith, 1979; Lin et al., 1998). U/Th ages from two of the salt beds—S6, 23.7 ± 0.7 ka, and S7, 22.6 ± 1.8 ka—are in agreement with the timing of Mono’s Big Low (Figure 4.27, Lin et al., 1998). Thus the similarity in the timing of dryness in the Mono and Searles Basins supports the idea that Last Glacial Maximum dryness was a Great Basin-wide phenomenon.

Along with the evidence of drying in the Searles and Mono Basins, there is evidence to suggest the plausibility of similar drying in the Lake Lahontan system during the LGM, too. Lake Lahontan reflects the integration of multiple, individual lakes from basins across northeastern California and western Nevada (Russell, 1885). In the Pyramid Lake sub-basin of the Lake Lahontan system (Figure 4.1), there is evidence to suggest that the lake had fallen below present conditions between the deposition of two tephra: the Trego Hot Springs and Wono Beds (Benson et al., 1997). The geochronology from this study suggests that the time encompassing the deposition of these two tephra is between 31.2 ± 0.2 ka and 27.5 ± 0.3 ka (Benson et al., 1997). Both of these time intervals overlap with the estimated limits of the Last Glacial Maximum. Thus it is a plausible hypothesis that the extreme low stand of Pyramid Lake occurred during the LGM. Furthermore, if the low stand occurred between 35-24 ka, it would be coincident with one of the five salt layers, S2-S6, in the Searles Basin. And it would, too, overlap with interpreted age of Mono’s Big Low (Figure 4.27).

The robust geologic data in support of LGM low stands in the Great Basin notwithstanding, the timing of the low stands are not sufficiently known to precisely test questions as to their synchrony. But the most conservative interpretations of their ages suggest that the hypothesis for simultaneous Great Basin low stands during the LGM is plausible. For
instance, the Big Low’s age estimates agree with the ages of S6 and S7 in the Searles Basin. But the precision of the ages also allows for the Big Low to instead correlate with the mud layer that is found between S6 and S7. Furthermore, the uncertainty on the ages of the Trego Hot Springs and Wono Beds preclude a confident determination of the age of the Pyramid Lake low stand. There is, therefore, a demand for high-precision geochronology on the sediments that constrain the low stands as well as the tephra that bracket them (if there are any). Without a robust chronology, questions as to regional synchronicities in dryness are equivocal.

4.12.3 Implications of LGM drying on the dynamics of ancient climate. Since the late 19th century, it is well understood that great lakes once occupied basins across the western US (Russell, 1885, 1889, 1895; Gilbert, 1890). And for nearly the last century, researchers have used a single hypothesis to explain the existence of these lakes (Antevs, 1952). This hypothesis advances that the temporal stability and anomalous cooling associated with the Laurentide Ice Sheet produced a persistent and strong high-pressure system about its surface (Figure 4.28). The permanence of this anomalous, high-pressure system forced a deepening of the Aleutian Low, a low-pressure system that is anchored in the North Pacific. This resulted in a southward shift in the mid-latitude jet that endured throughout the year, producing perennial wetness as opposed to the seasonal nature of the present water year. Ultimately this hypothesis predicts that lake high stands are coincident with the glacial maximum.

Because the evidence from Mono Lake (along with the evidence from Lake Lahontan and Searles Lake) for LGM drying contradicts the aforementioned hypothesis, I reason that perennial westerly storms did not characterize the Last Glacial Maximum. Nor do I find the assertion that
the Big Low may have been caused by storm tracks shifted even further to the south—say, because of an even greater deepening of the Aleutian Low at the glacial maximum—to be reasonable: for Lake Bonneville, a lake in the northeastern Great Basin, which should be defined by epic dryness by this hypothesis, was broadly transgressing during the time of the Big Low (Oviatt et al., 2015). I thus find the presence of the Laurentide Ice Sheet as the sole agent forcing western US lake fluctuation to be untenable. What, then, could the forcing mechanism be?

At present, the data is not sufficiently resolved to understand the climate dynamics that caused the epic low stands of Great Basin lakes during the Last Glacial Maximum. The most significant ingredient missing is high-precision dates on the lake low stands across the Great Basin. The present geochronologic data are only sufficient to show that they occurred during the coldest chapter of the last glacial cycle. In addition, the precision of the data, which are up to several kyr, do not allow for a test of regional synchronicity. Furthermore, the data do not allow a comparison to global hydroclimatic patterns, which are known to fluctuate at millennial and centurial scales (e.g., North Atlantic Dansgaard-Oeschger cycles, Wang et al., 2001).

4.1.3 Conclusion

Both the timing and extreme character of the Big Low challenge long-held conjectures of western US paleoclimate: that cool periods are wet and warm periods are dry. However, I am confident that my interpretation for an extreme low stand of Mono Lake during the Last Glacial Maximum is robust for the reasons detailed in this chapter: because of the fluvial gravel filling a cut at the “Big Low” type-section; because of the independent stratigraphic sequences that corroborate the Big Low findings; and because of the comprehensive sequence that provides
evidence for an apparently 120- to 150-vertical-meter oscillation of the lake between the deposition of Ash 5 and Ash 4—the lowest elevation of which occurred sometime between ~20.5 and 24.4 ka. In addition, the new dates I have presented in this study that bound the Big Low are consistent with the stratigraphy, and they deviate little from previously published chronologic constraints on the Wilson Creek strata. Furthermore, evidence from the Lake Lahontan system and Searles Lake corroborate the data constraining the Big Low, which implies that severe drying during the Last Glacial Maximum was a regional—not local—phenomenon.

References


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Company.


Table 4.1: Carbonate U/Th data

<table>
<thead>
<tr>
<th>LabID</th>
<th>Location</th>
<th>Elevation (meters)</th>
<th>238U (ppb)</th>
<th>232Th (ppt)</th>
<th>234U (measured)a</th>
<th>230Th/238U (activity)</th>
<th>Age (yr) corrected</th>
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<td>UWC5-1A</td>
<td>Upper Wilson Creek</td>
<td>2030 ± 13</td>
<td>2014 ± 12</td>
<td>0.00018 ± 0.00005</td>
<td>0.00018 ± 0.00005</td>
<td>0.2051 ± 0.0005</td>
<td>20847 ± 59</td>
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<tr>
<td>UWC5-1B</td>
<td>Upper Wilson Creek</td>
<td>2030 ± 13</td>
<td>2014 ± 12</td>
<td>0.00018 ± 0.00005</td>
<td>0.00018 ± 0.00005</td>
<td>0.2022 ± 0.0008</td>
<td>20509 ± 90</td>
</tr>
<tr>
<td>S49</td>
<td>Big Low type-locality</td>
<td>1982 ± 1</td>
<td>2011 ± 12</td>
<td>0.00018 ± 0.00005</td>
<td>0.00018 ± 0.00005</td>
<td>0.2336 ± 0.0014</td>
<td>24379 ± 182</td>
</tr>
<tr>
<td>S549</td>
<td>Tufa Crags</td>
<td>1991 ± 1</td>
<td>2011 ± 12</td>
<td>0.00018 ± 0.00005</td>
<td>0.00018 ± 0.00005</td>
<td>0.2556 ± 0.0008</td>
<td>2597 ± 105</td>
</tr>
</tbody>
</table>

Note: Corrected U ages assume a 230Th/234U atomic ratio of 10 ± 5 x 10^-7, which is similar to the present lake waters (Anderson et al., 1982).

U decay constant: $^{238}U = 1.55125 \times 10^{-10}$ yr$^{-1}$ (Jaffey et al., 1971) and $^{234}U = 2.82206 \times 10^{-6}$ (Cheng et al., 2013). Th decay constant: $^{230}Th = 9.1705 \times 10^{-6}$ (Cheng et al., 2013).

a: $^{234}U = \left(\frac{^{234}U}{^{238}U}\right)_{activity} \times 1000$

b: $^{234}U_{initial}$ is calculated based on the $^{230}Th$ age ($T$) -- $^{234}U_{initial} = ^{234}U_{measured} \times e^{234T}$

c: B.P. stands for "Before Present", Present is defined as 1950 A.D.
Figure 4.1. A Digital Elevation Model—taken from Reheis et al., (2014)—showing the western United States (see inset map) with the surface areas of its ancient lake high stands (light blue shapes) and locations of its inflowing rivers (dark blue lines). The overlain colors refer to mean annual precipitation (MAP) from 1971-2000 (PRISM). The geographic limits of the Great Basin are outlined in black. The lake basins referred to in this study are Lake Bonneville (LB) in the
eastern Great Basin, Searles Lake (labeled as SL next to a white circle) in the southwestern Great Basin, Mono Lake (labeled as ML next to a white circle) in the west-central Great Basin, and Lake Lahontan (LL) in the west-central Great Basin. Lake Lahontan’s two extant lakes are Walker Lake (labeled as WL next to white circle) and Pyramid Lake (next to white circle, PL). Other lakes shown on this map but not discussed in this study are Alvord (Al), Chewaucan (Ch), Columbus (Col), Coyote (Coy), Franklin (Fr), Fort Rock (FR), Newark (Ne), Manly (Ma), Smoke Creek Desert (SCD), Surprise (Su), Thompson (Th), Warner (Wa). The rivers labeled on this map are Amaragosa River (AR), Humboldt River (HR), Mojave River (MR), Owens River (OR). Major cities are labeled on the map as LA (Los Angeles), LV (Las Vegas), R (Reno), SLC (Salt Lake City), an W (Winnemucca).

Figure 4.2. NASA satellite image showing the Mono Basin and relevant locations (numbered 1-8) to this study. The yellow highlighted area shows location of the Mono-Inyo Craters. The blue, linear features are perennial streams that flow into the lake (1. Rush Creek; 2. Lee Vining Creek; 3. Post Office Creek; 4. Mill Creek; 5. Wilson Creek). The red box shows the approximate area shown in Figure 4.17. The Black Point cinder cone is feature #6. This study’s measured Lower Bridgeport Creek sections are feature #7. And the sedimentary sequence I observed at Cottonwood Creek is marked at feature #8. The “Tufa Crags” are found at feature #9.
Figure 4.3. Line plot show the relationship between elevation and predicted surface area of Mono Lake. I calculated the surface area-elevation relationship at a one-elevational-meter interval within a Geographic Information System using Digital Elevation Models after Gesch et al., (2002).
Figure 4.4. Photograph looking southwest at the western canyon wall of lower Wilson Creek. The photograph shows fluvial gravels overlain by Lajoie’s Wilson Creek Formation. The colored bars indicate the position of the numbered tephra, Ashes 19-1. Ash 1 is eroded from this section. It is, however, marked on the photograph. Ash 1 is found in the type-section, which is located upstream of this position.
Figure 4.5. Photograph looking southeast at the eastern canyon wall of Mill Creek. The red-circled areas show the approximate location of two sedimentary sequences described in this study.

Figure 4.6. Photograph showing the eastern canyon walls of Mill Creek of the “Big Low” type-locality. Labels S1, S2, S3 refer to units that comprise lacustrine silt. Labels G1 and G2 refer to units that comprise gravels (fluvial and lacustrine). And the dashed lines show the stratigraphic position that are concluded must represent disconformities. The lettered, white boxes show the approximate position of photographs explained in Figures 4.7-4.15.
Figure 4.7. (box “a” in Figure 4.6) Photograph showing S1 overlain by S3. A disconformity, which is shown by the white line, marks the two layers. Here S3 comprises cobble gravel, silt laminae, and massive to coarsening upward very fine to fine sands. The scale in the photo is a 28-cm-long ruler.
Figure 4.8. (box “b” in Figure 4.6) Photograph showing S1, G1-1, G1-2, G2, and S2. The contacts shown by the solid and dashed lines are disconformities. Ash 14 is partly shown in this photograph, but its lower contact is not clear. Ash 15 is also found in S1, but it lies below the observable field.
Figure 4.9. (box “c” in Figure 4.6) Photograph showing G1-1, G1-2, S2. Ash 2 is clearly observed. Ash 4 and 3 are not perceivable in this photo, but they are indeed found below Ash 2 (see Figure 4.10). A grey- and blue-colored sharpie pen is in the G1-1 strata for scale.
Figure 4.10. (box “d” in Figure 4.6) Photograph showing Ash 4, 3, and 2 in lacustrine silts (S3). The scale in the photo is a yellow-colored meter stick.
Figure 4.11. (box “e” in Figure 4.6) Photograph showing G1-1, G1-2, and the disconformity (the white line) that is marked between the two.
Figure 4.12. (box “f” in Figure 4.6) Photograph showing the gravel units that compose G1--1, 2, 3, and 4—and S3, the lacustrine silt that contains Ashes 4, 3, 2, and 1. The white lines show the contacts between the unit. I interpret each contact to be a disconformity except for the contact between G1-3 and G1-4 and between G1-3 and G1-2. And I infer that this contact is depositional. The red-lined box labeled “g” shows the position of box “g” in Figure 4.6. I use a white-colored, 28-cm-long rule, which is in the area highlighted by the red box, for scale.
Figure 4.13. (box “g: in Figure 4.6) Photograph showing a closer view of G1-2, G1-3, and G1-4. G1-2 and G1-4 are fluvial cobble gravel. G1-3 is a littoral embankment. The contacts between the G1 units, which are depositional, are marked by white lines. The dashed white line between G1-3 and G1-4 shows the approximate location of where the contact should be. The G1-3 deposit obscures its actual position. The result of the U/Th analyses measured on calcite cementing the G1-3 conglomerate (S49, see Table 4.1) is shown on the figure.
Figure 4.14. (box “h” in Figure 4.6) Photograph showing the silt block, S2, that was deposited within G1-1 and overlain by G1-2. S2 is strongly convoluted. I found Ash 7, Ash 6, and Ash 5 in S2; however, Ash 7 is the only clearly recognizable tephra in this photograph. Ash 6 and Ash 5, which measure only a few-millimeters-thick cannot be observed at this distance.
Figure 4.15. (box “i” in Figure 4.6) Photograph showing Ash 5 and Ash 6 in S2. I reason the double layer of Ash 6 is a product of prior deformation that resulted in convoluted stratigraphy. This is consistent with observations I made of the Ash 7, 6, and 5 sequence in sedimentary exposures that are no more than 100 m up-stream of this location (see section 4.4).
Figure 4.16. Photograph showing the wave-cut disconformity (red-dashed line) cross-cutting lacustrine silts that contain Ashes 5-7. And it shows the littoral sands that overlie this disconformity. Lacustrine silts overlie the littoral sands.
Figure 4.17. Photograph showing Mill and Wilson Creek and the northern half of Black Point. The location marked as “1” is the “Big Low” type-locality, which is show in Figures 4.6-4.15. The location marked as “2” is the location of the observed sedimentary section from the type-locality of the Wilson Creek Formation (Figures 4.18-4.20). The location marked as “3” shows the position of the described sedimentary section from Upper Wilson Creek (Figures 4.21-4.23).
Figure 4.18. Photograph showing one variant of the Big Low sands (Sd1) that I find overlying lacustrine silt containing Ashes 19-5 (S1) and Ashes 4-2 (S2). The sand bed in this photograph is approximately two-centimeters-thick.
Figure 4.19.: Photograph showing the western canyon wall of Lower Wilson Creek and the upper half of Lajoie’s Wilson Creek Formation type-locality. Two lacustrine silt units, S1 and S2, are bisected by Sd1, a well sorted fine to medium sand layer. I found Ash 5, 6, and 7 in S1. And I found Ash 2, 3, and 4 in S2. Ash 1 is not found at this location because of a younger, Holocene lake rise that eroded it from the section along with a portion of Ash 2. Ash 6 is too thin to be observed in this photo. I reason that Sd1 represents littoral deposition of sands during the lake’s rise following the Big Low. This, then, requires that the contact between Sd1 and S1 to be a disconformity (indicated here by the red-dashed line)—more specifically, an erosional surface as a result of wave planation. The contact between Sd1 and S2 is depositional. The crude north-dipping orientation of the lower half of S1 shows the deformed section of S2.
Figure 4.20. Photograph showing the Big Low sands (Sd1) between lacustrine silts containing Ashes 14-5 (S1) and Ashes 4-1 (S2). Sd1 is thicker at this location than in the prior photograph. It is also more strongly deformed.
Figure 4.21. Photographs “a” and “b” show the location (red circle) on the tufa mound that I sampled the calcite from in the field. Photograph “c” highlights the crystal (red dot) that were analyzed using the U/Th method. The resultant data is shown, too (UWC5-1A and UWC5-1B; see Table 4.1).
Figure 4.22. Photograph showing the stratigraphic context of the 20.5 ka tufa mound. Dr. Xianfeng Wang is pointing here to where we found Ash 4 overlying the sampled tufa mound.
Figure 4.23. Photograph showing Ashes 4-2 in lacustrine silts that abut the 20.5 ka tufa mound.
Figure 4.24. Photograph showing the north gully wall of lower Bridgeport Creek. I grouped the deposit of lacustrine silt shown here into three units: S1, which contains Ashes 8-5; S2, which contains no tephra; and S3, which contains Ashes 4-2. Ashes 3 and 2 are not seen in this photograph. S2 is unique. It is exceptionally rich in carbonate. And it contains calcareous laminae, which manifest as protuberant ridges. Because I infer that the expression of the “Big Low” must be between Ash 5 and 4, I attribute our observations of S2 to the low lake conditions. I have not determined a reasonable explanation for S2’s character. But I find no evidence of subaerial exposure in this sequence, which is measured at 1,955 m.
Figure 4.25. Photograph looking north towards the Tufa Crags. The buff-white-colored outcrops are lacustrine silts that contain Wilson Creek Formation tephra.
Figure 4.26. Photograph showing the U/Th-dated calcite coatings that were sampled from the Tufa Crags. The brown-colored material is the physio-chemical tufa that composes the structure of the Tufa Crags. The white-colored material coating the physico-chemical tufa is what I dated using the U/Th method. Note the speleothem-like structure of the coating adjacent to S549.
Figure 4.27. Plot showing the dates constraining the Last Glacial Maximum dry episodes from Mono Lake, Searles Lake, and Pyramid Lake. The black circles are carbonate U/Th dates from this study. Blue squares are U/Th dates on salt layers from Searles Lake (Lin et al., 1998). And the orange-highlighted area represents the time that occurred between the deposition of the Trego Hot Springs and Wono Tephra Beds (Benson et al., 1997). Thus the extreme low stand of Pyramid Lake occurred during the time highlighted in orange. Salt layers S7 and S6 from the Searles Basin show the best correspondence to the Big Low’s time constraints. The Pyramid Lake low stand is demonstrably older than the Big Low, but it is coincident with the S5 salt layer in the Searles Basin. With the present precision of the geochronology, the data show that some of the recorded Last Glacial Maximum low stands of Great Basin lakes are in agreement with each other.
Figure 4.28. A cartoon from Putnam (2016) showing the general hydroclimate conditions thought to have defined the Last Glacial Maximum. During this time, there was an anomalous high-pressure system associated with the North American ice-sheet, which comprises two parts: the Laurentide Ice Sheet (LIS); and Cordilleran Ice Sheet (CIS). The perennial nature of the ice sheet perturbed atmospheric circulation, causing the mid-latitude jet stream (blue line and arrow) to divert to the south. This caused lakes to rise in the Great Basin of the western US.
Appendix

Introduction

This appendix presents sedimentary, geomorphic, and chronologic evidence from the Mono Basin (Figure A.1) on the fluctuations of Mono Lake during four time intervals: 25-20.5 ka (Part 1); 20.5-13.8 ka (Part 2); 13-12 ka (Part 3), and 11-9 ka (Part 4). Most of my interpretations depend on all three evidential components used in conjunction with one another. This multipronged approach lends confidence as to the accuracy of my interpretations. Some interpretations are, however, based on evidence in the form of either sedimentary sequences, geomorphic surfaces, or dated material. These limitations notwithstanding, singular data are useful reference points on the limit of the lake’s elevation.

Part 5 of this appendix compares the Mono Lake fluctuation record from this study to the most recent of past interpretations of Mono Lake’s Late Pleistocene and early-to mid-Holocene fluctuation record (Benson et al., 1990).

Stratigraphy and Geomorphology

The Mono Basin’s Late Pleistocene lithostratigraphic unit is the Wilson Creek Formation (Lajoie, 1968). It is exposed in road-, wave-, and stream-cuts as well as deflation hollows. The
elevational range of these exposures are 1,955-2,155 m. Its type-locality is along the lower tracts of Wilson Creek (Figure A.2). There, the Wilson Creek Formation is defined by a six-meter-thick sequence of lacustrine silts, clays, sands, and 19 intercalated tephra. The tephra are numbered by their reverse emplacement: Ash 1 is the youngest; Ash 19 is the oldest.

Wilson Creek tephra are hugely important chronohorizons. Lajoie used them to piece together a relative lake fluctuation from isolated sedimentary exposures across the basin (Figure A.3). His interpretations of the sedimentary sequences were largely based on fundamental principles of sedimentation. But they were not interpreted in the context of sequence stratigraphy. And hence Lajoie’s study did not fully resolve the many disconformities preserved in the sedimentary exposures of Wilson Creek age. A model of lake fluctuation that incorporates sequence stratigraphy is needed for this study.

The methodology I employ to investigate Mono Lake’s stratigraphic record follows Stine (1987)'s principles of deltaic sedimentation in an endorheic lake basin. This model fuses geomorphology and stratigraphy, which allows for a wholesome examination of the lake’s ancient fluctuations. The stratal relationships predicted by the Stine model were used to constrain the late Holocene fluctuations of Mono Lake. I, too, observed nearly the same—and, in many cases, precisely the same—stratal relationships in the sedimentary sequences of Wilson Creek age. This suggests that the Stine model provides a reliable framework to deduce the lake’s late Pleistocene levels.
Chronology

Reliable geochronology is the greatest impediment to the determination of Mono Lake’s Late Pleistocene fluctuation record (Lajoie, 1968; Benson et al., 1990, 1998; Kent et al., 2002; Benson et al., 2003; Hajdas et al., 2004; Zimmerman et al., 2006, 2011, 2012; Vazquez et al., 2012). Although its sedimentary record includes many volcanic ash layers that can be dated using high-precision analytical techniques (\(^{40}\)Ar/\(^{39}\)Ar, for example, Kent et al., 2002), they are too infrequent to resolve the lake’s rapid fluctuations. Therefore a chronometer that can date other material that is commonly found in the lake is needed.

Radiocarbon dating is a common tool used for dating lake fluctuations of the last 30 ka. Terrestrial macrofossils are the most reliable material for this analytical method; however, this material is rare in sedimentary exposures of the Mono Basin. Tufa, in comparison, abounds in the Mono Basin.

Biotic and abiotic processes form tufa (Riding, 2000; Gierlowski-Kordesch, 2010). The size and fabric of this calcareous deposit varies in the Mono Basin (Russell, 1889; Dunn, 1953; Lajoie, 1968). These deposits, which range from centimeter- to meter-scale deposits, form as a result of lake- and spring-water mixing (Dunn, 1953; Figure A.4). The groundwater provides the calcium (Zimmerman et al., 2006), and the lake- and spring-water both provide carbonate in differing proportions (Benson et al., 1990). This physico-chemical process produces distinct macroscopic textures—namely, calcite that is porous and friable or brittle (Russell, 1889; Dunn, 1953; Scholl and Taft, 1964). Some fabrics occur in limited stratigraphic intervals. Thinolite, a calcite pseudomorph after ikaiite (Bishoff et al., 1993a), for instance, is exclusive to sediments that are younger than the Wilson Creek Formation (Figure A.5-A.6). And because ikaiite only
forms in near-freezing conditions (Bischoff et al. 1993b), its stratigraphic restriction is evidence for a unique period of anomalously cool water temperatures.

Photosynthetic algae are present in Mono Lake (Scholl and Taft, 1964). And it has been observed that they are intimately associated with a distinct fabric of tufa (Scholl and Taft, 1964; Scott Stine, personal communication) (Figure A.7-A.8). The tufa they form varies in thickness, but it is typically on the order of millimeters or centimeters and rarely decimeters. The key attribute of this fabric is that it is relatively denser than the physico-chemical tufa. It forms as planar to domal structures with laminae. Their colors are dominantly white to beige. Green- to blue-green-colored highlights are infrequent. Samples that appear to be detritus free are found coating tufa (Figure A.8, photo a). But those found in association with littoral sands or silts are frequently dirty—that is, varying proportions of detritus are found in them (Figure A.9); however, some coatings on larger clasts (cobbles, for instance) are relatively cleaner. These macroscopic textures and the stratigraphic context they are found in are similar to modern and ancient lake carbonates that are interpreted to be algal tufa (Gierlowski-Kordesh, 2010).

Lacustrine tufa is most commonly dated by the radiocarbon method. But $^{14}$C dating of this material (as well as other calcareous lacustrine macrofossils like ostracods) in the Mono Basin is unreliable for the following reasons: because ground- and lake-waters contain unknown dead carbon (Benson et al, 1990), resulting in dates that are likely too old; and because the magnitude of modern carbon contamination, which causes dates to be too young, is uncertain (Thompson et al., 1986; Kent et al., 2002; Zimmerman et al., 2006).

U/Th dating is an alternative methodology to date lacustrine tufa (Lin et al., 1996; Lin et al., 1998). But elevated initial Th makes this approach challenging (Anderson et al., 1982; Zimmerman et al., 2006). Isochron regressions are sometimes used to correct for this
unsupported intermediate daughter (e.g., Lin et al., 1996; Ibarra et al., 2014). New studies have shown success in dating dense abiotic tufa (McGee et al., 2012; Steponaitis et al., 2016) with minimal initial Th. Similar high-precision U/Th dates (~0.3% uncertainty, two sigma) on biotic and abiotic tufa are reported from the Mono Basin (Wang et al., 2011).

Algal tufa are the principal underpinning for this study’s chronology. Their utility is two-fold: their limited initial Th yield reliable and high-precision ages, which allows for a centurial resolution of the lake’s fluctuations; and their dates provide a lake level constraint because they only form in the photic zone. I follow the methodology used by the Wang et al., (2011) for dating tufa by the U/Th method for this study. And the result of this undertaking is the first century-scale lake fluctuation record from a continuously-closed basin during the period encompassing the late glacial period to the early Holocene.

**Elevational uncertainty on algal carbonates**

Six glacial streams were active in the Mono Basin during the last glacial period (Russell, 1989) (Figure A.10). Comminution of bedrock and debris by glaciers elevates the concentration of suspended material in proglacial lakes (Haldorsen, 1981). This influx of suspended material to the lake affects the depth of the photic zone, which is inversely related to turbidity (Edmunson and Koenings, 1986). Thus glacial lakes, which are defined by greater turbidity relative to clear-water lakes, have photic depths that are shallower than clear-water lakes (Edmunson and Koenings, 1986). In glacial lakes, it is estimated that the photic zone is no deeper than four to six meters (Edmunson and Koenings, 1986). Thus during the time that the Mono Basin was glaciated, photic algal communities would have likely resided in the top six meters of the lake’s water
column. And by this understanding, I reason that U/Th dates on algal carbonates reflect an elevation within five meters of the lake’s surface for the duration of the Mono Basin’s glaciation.

It is demonstrated that glacial streams were active in the Mono Basin during the last glacial period (Russell, 1889). Their retreat from their maxima occurred ~19 ka (Rood et al., 2011). It is unknown how long these local glaciers persisted. But it is estimated that the Sierra Nevada was deglaciated by ~13 ka (Clark and Gillespie, 1997). I reason that turbidity, too, would decrease significantly in Mono Lake thereafter due to the absence of glacial silt. If the lake’s turbidity after 13 ka was similar to clear-water lakes today, then the depth of the photic zone could be greater than 20 m during the Younger Dryas and early Holocene. Thus the vertical uncertainty on the U/Th dates from this time precludes a direct estimation of lake level from ancient algal carbonates. Therefore, I interpret the dates on algal carbonates younger than 13 ka as minimum elevational constraints. And in the context of this study, I argue that the vertical uncertainty associated with these data prior to 13 ka is +5 m. Because this uncertainty is relatively small compared to the century-scale lake fluctuations resolved here (>30 m), did not mark this uncertainty on the data plotted in the figures.

On uncertainties of paleoelevations as a result of tectonic deformation

The Great Basin of the western US is a region of active continental rifting (Stewart, 1971). Its westernmost margin is the Sierra Nevada (Hammond and Thatcher, 2004). The Mono Basin is a partially-filled half-graben that abuts the eastern Sierra Nevada (Pakiser, 1960). The Sierran frontal fault, marks their boundary. Seismic and drill-hole data show no major faults that cut the half-graben (Pakiser, 1960).
Down-dropped morainal embankments that protrude from the Sierra’s U-shaped canyons provide evidence for fault activity (Russell, 1889). The most striking evidence of faulting is the ∼20-vertical-meter offset across the last glacial maximum moraine of Lundy Canyon (Bursik and Sieh, 1989; Rood et al., 2011b). \(^{10}\)Be surface exposure ages on this moraine imply that it is ∼19 ka (Rood et al., 2011a).

Many of the data I use to support my lake fluctuation record are from the northwestern Mono Basin, which is proximal to the fault segment that displaced the Lundy Canyon moraines. Present elevations of ancient sedimentary deposits need to be corrected for tilting associated with the displacement accommodated by the Sierran frontal fault. But the scope of this dissertation did not allow for this uncertainty to be fully evaluated. And since this uncertainty was not resolved in past studies, I chose not to implement tectonic corrections in this study to avoid confusion.

**Part 1. Evidence for an epic low stand of the lake and its rise thereafter during the late-glacial period (25-20.5 ka)**

**The Big Low type-locality.** Two nested cut-and-fill sequences that truncate thinly-laminated silts occur along the eastern channel walls of Mill Creek (Figures A.1, A.11, A.12). There are no published descriptions of this outcrop. I describe it from observations I made across four-vertical meters and 35-lateral meters of exposure.

The first cut-and-fill sequence (G1) is divided into four units (1-4) (Figure A.18): a lower composite unit comprising well sorted, planar and wavy sands (G1-1) and massive, poorly sorted
gravel that ranges up to cobbles in size (G1-2) (Figure A.17); a medial unit comprising very well sorted, cross-laminated, pebble conglomerate (G1-3) (Figure A.19); and an upper unit of moderately sorted sands and granules and poorly-sorted gravel up to cobbles in size (G1-4) (Figure A.18). The lower unit is underlain by thinly-laminated silts and clayey-silts of the Wilson Creek Formation (S1) and Ashes 15-8 (Figure A.14). The contact below the sand and gravel sequence is erosional. Intraformational clasts of Ash 8 and an intraformational block of highly-deformed lacustrine silt (S2) containing Ashes 7-5 are found in the lower fill unit (Figures A.20-A.21). I sampled calcite crystals that cemented the conglomerate of the medial unit (G1-3) for U/Th analysis.

The second cut-and-fill sequence truncates the first. Its basal contact is curvilinear (Figure A.12-A.14). Like the first sequence, it incises into the thinly-laminated silts that contain Ashes 15-8, too (Figures A.14). The strata that compose the second cut-and-fill sequence are threefold. The lowest exposure is comprised of poorly sorted gravel up to cobble in size (G2, Figure A.14). The G2 gravel abuts similar strata that composes the first fill sequence (Figure A.14). Overlying the G2 gravel is a two-part sequence that abuts and overlies the first cut-and-fill deposit (Figure A.13-A.14). This bipartite sequence comprises thinly- to thickly-laminated and variably deformed silt and very well sorted sands, which are in turn overlain by laminated silts containing Ashes 4-1 (S3, Figure A.15). Poorly sorted gravel up to cobble in size overlies Ash 1.

These stratigraphic data constitute the lowest elevation of recognized evidence for riverine strata in the Wilson Creek Formation. The observations described above require that the lake regressed to below 1,982 m following the deposition of Ash 5. During the time encompassing this “Big Low” low stand, a stream incised through Wilson Creek lacustrine
strata. This incision must have formed a canyon. I interpret the curvilinear contact between the two cut-and-fill sequences to be the surface of this ancient canyon. The lacustrine silts that overlie this paleosurface—and the tephra they contain—require that the lake transgressed from the Big Low to some elevation above this deposit prior to the deposition of Ash 4. I, therefore, deduce that the Big Low occurred between the deposition of Ash 5 and Ash 4.

The cementation of the littoral embankment with calcite is enigmatic because I do not find it in any other deposit. Because the calcite cement is exclusive to the littoral embankment, and because the strata are not different than those below or above it, I infer that the calcite likely formed as the littoral embankment was being deposited. By this reasoning, the U/Th datum on the calcite would closely constrain the timing of the littoral embankment’s deposition. And because the second cut-and-fill sequence cross-cuts the embankment, the U/Th datum is a maximum age for when the lake transgressed from the Big Low.

**Observations from upstream of the Big Low type-locality.** Approximately 200 m upstream of the Big Low locality, and also along the eastern channel walls of Mill Creek, the full set of Wilson Creek Formation tephra are found in thinly- to thickly-laminated silts that are underlain by poorly sorted, pre-Wilson Creek gravel (Figure A.12). The elevation of the Wilson Creek Formation here is 1,986 to 1,993 m. Within the interval between Ash 5 and 4, I found a deposit that I infer to represent the expression of the Big Low (Figure A.21). This deposit comprises very well sorted sands that are underlain by a horizon of pebbles and cobbles. The basal contact of this deposit, which I measured to be ~1,991 m, is sharp. The contact cross-cuts a convoluted interval of silt that contains Ashes 7-5.
I interpret the poorly sorted cobble gravel as fluvial gravel, the thinly-laminated silts as lacustrine silts, and the very well sorted sands as littoral sands and the cobble and pebble clasts as lag deposits. I, therefore, conclude that the contact between the littoral sands and the lacustrine silt it overlies is a wave-cut disconformity. I interpret that the lake fell below the 1,991-m disconformity within the time encompassing the deposition of Ashes 5-4.

**Wilson Creek type-locality.** Because the Big Low type-locality requires the lake to have fallen below 1,982 m, there must be a correlative disconformity between Ashes 5 and 4 in the Wilson Creek type locality: for the sediments exposed along lower Wilson Creek are as low as 1,970 m and as high as 1,999 m. Previously published observations document the presence of deformed silts and sand lenses between Ashes 5 and 4 (Figures A.2 and A.23; Lajoie, 1968; Benson et al., 1998).

The sands found between Ashes 5 and 4 are very well sorted (Figure A.24). Their lowest exposure is 1,975 m; their highest exposure is 1,989 m. They are underlain by planar parallel, thinly- to thickly-laminated silts that are intercalated with Ashes 19-5. And they are immediately overlain by a one-meter-thick interval of up-thrusted, thinly-laminated silts (Figures A.25-A.26). The deformed laminae dip moderately to the north. These deformed silts are overlain by planar parallel, thinly-laminated silts intercalated with Ashes 4-1 (Figures A.25-A.26).

Lajoie considered the sands to reflect deposition by sublacustrine mass wasting. But a different study suspected that they represented a disconformity between Ashes 5 and 4 (Benson et al., 1998). Though it is conceivable that these sands are sublacustrine mass wasting deposits, the most parsimonious explanation that is consistent with the other stratigraphic data along Mill
Creek is that the sands are littoral sands that were deposited on a surface of wave planation. The low-angle, lakeward dip of this wave-cut disconformity is further consistent with this interpretation. If correct, this interpretation would extend the elevational limit of the Big Low from 1,982 m—which I concluded from the Big Low type-section—to 1,975 m.

**Upper Wilson Creek.** Where Cemetery Road crosses Wilson Creek (Figures A.1, A.12, and A.27) and approximately two kilometers upstream of our observed sections at the Wilson Creek Formation type-locality, Lajoie (1968) reported on an eight-meter-thick sequence (labeled as II-E) he measured that comprises well sorted sands and gravel that are overlain by silts containing Ashes 4-1. Lajoie, too, found physico-chemical tufa mounds rooted in the sands and gravel. Here I report on the same eight-meter-thick sequence with new observations.

My observations of Lajoie’s II-E show a sedimentary sequence comprising three successive units. The lower unit comprises tabular, cross-laminated, moderately- to well-sorted, granule gravel and sands that contain abundant rounded pumice lapilli. It is overlain by the middle unit, which comprises six meters of unconsolidated, very well sorted sands that interfinger with poorly sorted pebble conglomerate and poorly sorted pebble gravel. A meter-and-half of thickly- to thinly-laminated silts intercalated with Ashes 4-1 compose the upper unit.

I found spring tufa mounds rooted in the basal contact of the third unit. I sampled two physico-chemical calcite crystals from one of these tufa mounds for U/Th analysis (Figure A.27). This tufa mound is overlain by Ashes 4-2 (Figures A.27-A.28). Ash 1 is missing from the sedimentary sequence overlying this tufa mound; however, it is clear that Ash 1 was eroded.
away because I observed it in adjacent sedimentary sequences. I dated the calcite I sampled using
U/Th analysis.

I interpret the tabular cross-bedded sand and granule gravel as a littoral embankment, the
well sorted sands from the second unit as littoral sands, the pebble gravel and conglomerate that
interfinger with the littoral sands as fluvial strata, and the laminated silts as hemipelagic
lacustrine silts.

I am unable to identify the unknown tephra in the first unit at this time, but I reason that
the abundance of pumice lapilli are characteristics consistent with Ash 7 or Ash 11—the only
two Wilson Creek tephra found in the northwestern Mono Basin that contain pumice lapilli
(Lajoie, 1968).

I deduce that this tripartite sedimentary sequence reflects a period of lake regression
followed by a period of lake transgression. I interpret the littoral embankment to be deposited at
lake level. And I interpret the overlying interfingering littoral and fluvial strata to indicate that
the lake was oscillating at their approximate elevation. And because the deposition of fluvial
strata is restricted to embayed delta trenches during lake transgressions (Stine, 1987), and
because the sediments that compose this sequence were not deposited in an incised canyon, I
reason that this sedimentary sequence reflects a period of lake regression and not lake
transgression. I cannot determine the specific time of this lake regression; however, I do know
that it must be prior to the deposition of Ash 4. I infer that the change from littoral and fluvial
sediments to lacustrine silts indicates that the lake rose after they were deposited. And because
Ash 4 overlies the change in strata from lake regressive to lake transgressive deposits, I reason
that the lake rise initiated prior to the deposition of Ash 4. And based on the evidence presented
thus far on the Big Low—that the lake rose from at least 1,975 m during the interval between
Ashes 5 and 4--I conclude that there is a wave-cut disconformity present in this sedimentary sequence. I argue that because the strata below the silts were deposited during a time of lake regression, the most plausible horizon to mark the disconformity would be the contact between lacustrine silts and the littoral sands and gravel. And because the tufa mounds are rooted at this contact, their formation, which I infer was a result of groundwater degassing or lake- and ground-water mixing, is likely contemporaneous with the lake’s rise.

Although I cannot directly correlate each depositional unit in the context of the Big Low, the stratigraphy is consistent with my other findings—namely, that the lake fluctuated to a low stand following the deposition of Ash 5 and rose again prior to the deposition of Ash 4. Thus the time when the lake rose from the Big Low to 2,030 m can be constrained by a U/Th date on the tufa mound that was overlain by Ash 4.

**Lower Bridgeport Creek.** The lowest elevation outcrop of the Wilson Creek Formation that I have recognized measures from 1,955 to 1,960 m. It is located approximately one km from the present shore of Mono Lake (~1,945 m) along the incised walls of the lowermost reaches of Bridgeport Creek (Figure A.1). The sedimentary sequences there that contain Ashes 15 through 2 are cut and filled with Holocene deposits (Scott Stine, personal communication). At the specific sedimentary sequence described here, I find a continuous sequence of planar parallel, thinly- to thickly-laminated silt intercalated with Ashes 8-3 (Figure A.30). Between Ashes 5 and 4, many of the laminae are marked by calcareous horizons. I found no sand or gravel deposits and no obvious disconformity in this interval.
The simplest explanation of these observations is that this exposure was continuously flooded by the lake throughout the interval encompassing the deposition of Ashes 15-2. I infer, therefore, that the lake did not fall as low as 1,955 m during the Big Low.

**Israel Russell’s Tufa Crags.** A 17-meter-high tufa deposit, termed the Tufa Crags (Russell, 1989), is exposed along a west-facing, wave-cut cliff (Figure A.31). Thinly-laminated silts containing Ashes 15-1 abut the Tufa Crags. I also find silts within the gaps and interstices of the tufa. Within one of these interstices, I found two calcite deposits that were akin to those found in caves (Figure A.32). One deposit grew from the roof of the interstice, and its structure appeared to droop downwards from the roof. The second calcite, which was on the floor of the interstice, grew upwards like a stalagmite. Both deposits cut across Ash 5. I sampled the calcite deposit that formed as a stalagmite for U/Th analysis. Its elevation is 1,990 m.

I interpret the tufa mounds as sublacustrine spring deposits that formed prior to the deposition of the oldest tephra it is overlain by. I do not argue that the tufa was formed in one generation, and in fact, I reason that there were likely multiple episodes of tufa formation by their varying textures. Further study in this area could be fruitful for understanding ancient fluctuations of Mono Lake.

The carbonate deposits’ relationship to Ash 5 requires two pairs of observations and interpretations: that the calcite cross-cuts—and is therefore younger than--Ash 5; and that based on the macroscopic texture of the calcite, it formed in a subaerial, cave-like environment. By these interpretations, the lake was below 1,990 m when the calcite was precipitated. These findings are consistent with our elevational interpretations on the Big Low. Thus the U/Th age on
the calcite that cross-cuts Ash 5 provides two constraints: a minimum age for Ash 5; and a maximum age for the Big Low.

**Cottonwood Canyon.** A four-meter-thick sedimentary sequence exposed along the eastern canyon walls of Cottonwood Creek (Figure A.1) includes Ashes 8-5 in silts and clays. I measured the base of Ash 8 to be ~2,007 m. I found clays, coarse silts, and two beds of medium sands between Ashes 8 and 7. The lower bed is nine-cm-thick. It has granules at its base, and it fines upwards. The upper bed is two-cm-thick, and it has rounded pumice lapilli. Ashes 7-5 are intercalated in planar parallel, thickly- to thinly-laminated fine to coarse silts.

I use the Cottonwood Canyon sequence to place a lower limit on the elevational range of the lake during the time when Ash 5 was deposited. Based on this evidence from Cottonwood Canyon and other sedimentary sequences described here, I estimate the minimum elevational fluctuation of Mono Lake during the fall to the Big Low and the rise thereafter. I interpret the silts and clays as hemipelagic lacustrine sediments, both sandy beds as littoral deposits, and the granule gravel at the base of the lower sands to be a lag deposit. Thus I infer that the granules and sands in the lower bed are littoral sediments deposited onto a wave-cut disconformity. This requires that the lake fell to below ~2,007 m during the time interval encompassing Ashes 8 and 7. I do not, however, find evidence of a similar disconformity between Ashes 7-5. And I reason, therefore, that the lake did not fall below ~2,007 m during that time interval, Thus the data suggest that lake level was at least 2,007 m at the time Ash 5 was deposited.
**On the lower elevational limit of Ash 4.** In order to determine the elevational limit that the lake rose to after the Big Low and prior to the deposition of Ash 4, I measured the highest elevation of Ash 4 in lacustrine silts. I found Ash 4 in lacustrine silts along the canyon walls of Rush Creek and Bridgeport Creek at 2,052 m. Additional confidence in this elevational estimate is supported because these two localities are from opposing sides of the basin. Thus the lake rose from the Big Low—with an elevational range between 1,975 m and 1,957 m—to 2,052 m at the time Ash 4 was deposited.

**U/Th dating (Table A.1).** I present U/Th data from two calcite samples from the Upper Wilson Creek site (UWC5-1A and -1B, Figure A.27), two from the Tufa Crags site (S549 and S550, Figure A.32), and one from the Big Low site (GA-S49, Figure A.19). I selected calcite that was dense and white, and apparently free of detritus. I sonicated the samples in ultra-pure water. And I used a carbide dental tool to mill ~1 mg of powder from the samples. The uranium and thorium in the sample was separated and purified following a methodology similar to Edwards et al., (1987) and Cheng et al., (2000). The uranium and thorium was measured the uranium and thorium isotopes by inductively-coupled mass spectrometry using a Thermo-Scientific Neptune 2 at Earth Observatory of Singapore. I presume that the present lake water $^{230}\text{Th}/^{232}\text{Th}$, which is $10 \pm 5 \times 10^{-6}$ (Anderson et al., 1982), is a reasonable estimate of the correction for calcite deposits of Wilson Creek age. The results of the analytical experiment are listed in Table 3.1. Included in this table are analyses made prior to this study by Dr. Xianfeng Wang in Dr. Larry Edward’s lab at the University of Minnesota.
The U/Th analyses yielded dates for UWC5-1A and -1B of 20,207 ± 193 a and 20,463 ± 237 a, respectively. This implies that Ash 4 is younger than 20.5 ka. And it suggests that the lake rose up to 2,030 m from the Big Low by this time, too.

The U/Th experiments on S549 and S550 from the Tufa Crags yielded ages of 24,478 ± 329 a and 25,106 ± 537 a, respectively. Because both samples cross-cut Ash 5, I conclude that Ash 5 must be older than 25.1 ka. And because the calcite is meteoric in origin, I deduce that the lake must have been lower than the elevation of the calcite samples--1,990 m--as early as 25.1 ka and as late as 24.5 ka. And I conclude that the chronologic constraints imply that the lake fell to within 15 to 34 m of the Big Low by 25.1 ka.

Using the modern lake composition to correct for initial thorium, my U/Th age on the calcite sample that cross-cuts the littoral embankment is 22,219 ± 1493 a. Its uncorrected U/Th age, 24,379 ± 182 a, is a maximum age. The deposition of the calcite as well as the littoral embankment, therefore, can be no older than ~24.4 ± 0.2 ka. And the Big Low is thus younger than ~24.4 ka ± 0.2 ka. This estimate is consistent with the inference that I made with the ~25 ka and ~24.5 ka calcite from the Tufa Crags.

(Part 2) The continued transgression of the lake to its highest level and its regression from it (20.5-13.8 ka)

It has long been recognized that Mono Lake reached its highest lake level of the last glacial cycle after the glacial maximum. Israel Russell first made this interpretation (Russell, 1889). He noted lake terraces carved on the youngest morainal embankment protruding from Lundy Canyon. He inferred that the embankment was deposited during the last glacial period.
And he reasoned, therefore, that peak wetness post-dated glacial maximum. Russell (1889) correlated the highest elevation terrace there to the most prominent circum-basin terrace. Its elevation is 2,155 m. $^{10}$Be surface exposure dating demonstrates that this moraine was deposited ~19 ka (Rood et al., 2011a). Thus the lake high stand is younger than 19 ka.

My dates corroborate Russell’s interpretation: that glacial maximum preceded peak wetness. I show that lake reached its maximum ~16 ka. And I show that the lake regressed 140 m from its 2,155-m high stand within two intervals: 16-15 ka; and 14.1-14 ka. The majority of these constraints are from calcite U/Th dates on conglomerate or algal coatings on tufa mounds. The evidence is expanded below.

**Grand Gash (Figures A. 33-A.35).** A 900-m-long, NNW-SSW-trending artificial ditch exposes up to 25-vertical meters of Wilson Creek strata. Scott Stine dubbed this sedimentary exposure “Grand Gash”. I describe three sedimentary sequences there. These sequences are termed A, B, and C.

**Sequence A (Figures A.33-A.34).** The lowest elevation and most lakeward sequence is defined by poorly sorted gravel with sizes up to cobbles (G2, Figure A.33-34). I found the gravel exposed from 2,010 to 2,022 m. This gravel is overlain by one- to two-meter-high thinolite mounds. I interpret the cobble gravel as fluvial strata. This implies that the lake was lower than 2,010 m when they were deposited. And I infer that the presence of the thinolite mounds indicate sublacustrine spring activity. Thus the lake must have transgressed from below 2,010 m after the
gravel was deposited. The minimum elevation that the lake transgressed to is the crest of the thinolite mounds, ~2,024 m.

I collected two samples of dense, white, laminated calcite that coated the elongate, thinolite crystals that composed the structure of the mounds. Based on their similarity to modern or late Holocene examples, I interpreted the calcite as algal deposits. I determined the age of the algal calcite by U/Th analysis. One age is 13.07 ± 0.13 ka; the other is 12.82 ± 0.3 ka. I interpret these ages as minimum lake constraints. Thus these ages imply that the lake transgressed from an elevation below 2,010 m to 2,024 m by ~13.1 ka.

**Sequence B (Figures A.33-A.34)** The second sedimentary sequence of Grand Gash is up-cut of Sequence A. Sequence B comprises poorly sorted cobble gravel(G2) that is overlain by a package of finer sediments, silts and sands (S1, Figures A.33-A.34). These fine strata measure up to seven-meters-thick, and they comprise two parts: a lower half; and an upper half. The lower half comprises the following strata: massive, poorly sorted pebble gravel with intraformational clasts of laminated silt and well sorted sands; sandy pebble gravel that fines upwards or coarsens upwards; massive, well sorted sand beds that are convoluted; sand and silt beds that fine upwards; and sandy beds that coarsen upwards and have floating pebbles and granules. The upper half comprises cross-laminated sands and thinly-bedded to thinly-laminated rhythmic silts. The rhythmic silts are intercalated with Ashes 4, 3, and 2. I do not find Ash 1 in this sequence. I measure the elevation of the contact between the gravel and the silts to be ~2,021 m. And I measure the elevation of Ash 4 to be ~2,028 m.
I interpret the cobble gravel as fluvial gravel. It is the same fluvial gravel as exposed in Sequence A. And I reason that the upper half of the silt package represents lacustrine mass-wasting deposits (turbidites and debrites). Because I find no evidence to suggest the upper half’s rhythmic silts were deposited en masse, I conclude that they must be lacustrine hemipelagic deposits. And similar to the interpretation I made in Sequence A, I interpret the fluvial gravel to indicate that the lake was lower than their elevation at the time of their deposition. And because I find Ash 4 (and Ash 3 and Ash 2) in the lacustrine strata above the fluvial-lacustrine contact, I reason that the lake must have transgressed from a lower elevation sometime prior to Ash 4. Thus the contact between fluvial and lacustrine strata is a disconformity.

*Sequence C (Figures A.34-35).* The third sedimentary sequence is up-cut of Sequence B. It comprises 14-vertical meters of poorly sorted granule gravel beds and silty-sand beds (G2, Figures A.34-A.35), which broadly coarsen upwards and dip to the south. The base of the sequence is ~2,028 m. It is capped by an approximately one-and-half-meter-thick deposit of poorly sorted gravel that ranges in sizes from pebble up to cobble (Figure A.35). Crude south-dipping bedding is observed at its base. And clear south-dipping bedding is observed at its top. In some isolated zones, the gravel is a conglomerate. Its cement is calcite. The top of the exposure measures ~2,042 m.

I interpret the cobble and pebble gravel and conglomerate that terminates the sequence as fluvial deposits. And although I could not find direct evidence for the contact between a top-set and fore-set bed, I argue that the most reasonable interpretation for the 14-vertical meter, coarsening-upward sequence of sands and gravel is that it reflects riverine material deposited
into the lake by a prograding stream. There is no indication that this riverine material was deposited in a delta trench, which is what would be expected to occur had the lake been rising during the time the strata was being deposited (Stine, 1987). Thus I reason that this sequence must have been deposited during a recession of the lake.

I collected two samples of white, calcite crystals from the fluvial cobble conglomerate that terminates the sequence. I interpreted the calcite to represent a minimum age for the deposit’s deposition. I determined the age of the two calcite samples by U/Th analysis (Figure A.35). The two samples yielded similar ages of 19.03 ± 0.60 ka and 19.10 ± 0.43 ka. Thus the lake was at 2,042 m by ~19 ka.

Grand Gash Interpretation. The fluvial cobble gravel that I found between 2,011 and 2,024 m in Sequence A and B, which I term “G2”, is a distinct deposit relative to the fluvial gravel of Sequence C, which I term “G1”. I argue this because the G2 cobbles do not continue rangeward at the same elevation. Rather what I find in the rangeward continuation of the G2 cobbles in Sequence C are finer gravels, silts, and sands. This precludes the possibility that the G1 and G2 gravels are syndepositional units. The paleo-geomorphic relationship, therefore, indicates that the G2 gravels abut the lakeward termination of the G1 gravels. And because the sedimentary sequences of Grand Gash are not confined to the delta trench, the only conceivable method by which the G2 gravels were deposited must be during a lake regression. Thus the G1 gravels represent riverine material deposited into the lake as a subsidiary delta--presumably of the ancient Mill Creek. And the G2 gravels are riverine material deposited as the topset beds of a subsidiary delta that abutted the lakeward edge of the delta associated with the G1 gravels. This
scenario explains the nature of the contact between the unique G1 and G2 fluvial deposits. And it suggests that the lake regressed from an elevation of ~2,042 m to at least 2,011 m prior to the deposition of Ash 4. The ~19 ka calcite U/Th ages from the G1 gravels do not constrain its depositional age. Rather the U/Th dates are a minimum depositional age and, therefore, a minimum lake level constraint, too.

Ash 1 is missing from S1 in Sequence B. Its absence is baffling—namely, because I find Ashes 4, 3, 2, and 1 in a continuous sequence of lacustrine silts in widely dispersed exposures across the basin from ~1,980 to ~2,055 m. This, too, requires that Ashes 4-1 are missing from Sequence A and C. I conclude, therefore, that a disconformity must be present in these sequences. The surface of Grand Gash marks this disconformity. I interpret its low-angle profile to indicate the surface was formed by wave-planation during a lake transgression.

Because the thinolite mounds overlie the disconformity that concurs with the outcrop’s surface, and because this disconformity is younger than Ash 1, I deduce that the thinolite mounds must be younger than Ash 1, too. This deduction implies that the lake rose from some elevation below 2,010 m to at least the top of Sequence C, ~2,042 m, after the deposition of Ash 1: for Ash 1 is found in sedimentary sequences as high as 2,145 m. And after the lake had flooded the G2 gravels exposed in Sequence A, the thinolite mounds formed. The U/Th ages on the algal calcite that coat the thinolite date the minimum age of the lake rise to ~13.1 ka. This 13.1 ka datum is also a minimum depositional age for Ash 1.

**Eastern Sierra Nevada (Figures A.36-A.37).** I show that Mono Lake rose from the Big Low to 2,155 m during the time encompassing 20.5-16 ka. Five calcite U/Th dates support this
interpretation. One date, \( \sim 19.5 \text{ ka} \), is on calcite-cemented pebble conglomerate at 2,040 m (see section on Grand Gash for explanation). Three U/Th dates are on calcite from isopachous-cemented conglomerate from the eastern Sierra Nevada. These analyses resulted in the following data: a date of 18.8 ka at an elevation of 2,070 m (Figure A.36); a date of 16.6 ka at an elevation of 2,105 m; and a date of 15.92 ka at an elevation of 2,145 m (Figures A.37). These four U/Th dates are minimum elevational constraints for the lake.

The 15.92 ka date is on calcite that cements a bimodal conglomerate. Well sorted subangular pebbles and poorly sorted angular detritus with sizes up to boulder compose the two parts. I deduce the poorly sorted boulder gravel are lag deposits, which were presumably deposited as alluvial material prior to their flooding by the lake. And I reason that the well sorted pebbles indicate reworking by littoral currents. I found similar deposits as low as 2,135 m and as high as 2,145 m at this location. I could not trace the deposits to the 2,155-m high stand due to vegetation cover. Because the angular nature of the well sorted pebble gravel indicates minimal reworking, I reason that the lake’s stay between the 2,135 to 2,155 m was transient. By this reasoning, I surmise that the lake reached the high stand at \( \sim 16 \text{ ka} \).

**Black Point (Figure A.38).** I also measured the U/Th age of calcite that cross-cuts Black Point. The calcite that was sampled coated a fracture that cross-cut the upper part of Black Point, which is welded and palagonitized (Figure A.38). The sample’s elevation is 2,090 m. The sedimentary sequence that the fracture cross-cuts is continuous with the sequence that composes the edifice of Black Point. But because I cannot determine if this calcite was formed by sublacustrine or subaerial processes, I can only argue that it serves as a minimum depositional
age for Black Point and, by extension, Ash 2. My U/Th analysis of the calcite yielded an age of 17.3 ka. Black Point and Ash 2 are, therefore, older than 17.3 ka.

**Cowtrack Mountains (Figure A.39).** I used the U/Th method to date calcite that cemented rounded pebble conglomerate that abutted the Cowtrack Mountains (Figure A.39). The conglomerate’s elevation is 2,120 m. The rounded nature of the pebbles suggest that they were worked by littoral currents. Replicate U/Th analyses on a single aliquot yielded ages of 15.4 ka. These data are a minimum elevational constraint for the lake.

**Thinolite towers along Highway 167 and Goat Ranch Rd cutoff (Figures A.40-A.41).** I found tufa mounds near where Upper Wilson Creek and Goat Ranch Road cut-off intersect. Their elevations were ~2,075 m. Dense, laminated coatings on physico-chemical tufa are abundant at this site. They are domal to botrioidal in form. And they measure up to several cm in thickness, and their color ranges from white to pale-yellow. I inferred that they were algal because of their morphology. I also found calcite coatings on thinolite mounds at an elevation of ~2,075 m. I interpreted these coatings to be algal deposits, too. I dated a total of 13 separate calcite coatings that ranged from a few-mm- to several-cm-thick. I measured two to six separate aliquots in stratigraphic order from the thicker samples. In total, 29 different aliquots were measured. The U/Th analyses yielded dates that ranged from 15.1 ka to 14.1 ka. Dates from the thick samples encompassed 14.6 to 14.0 ka. The dates are replicated on samples from different tufa mounds. And because I interpreted the deposits as algal-mediated
calcite, I reason that the lake remained at 2,075 m during this time. The macroscopic textures of the thick samples, which suggest continuous algal growth, support this interpretation. The corroboration between the macroscopic and chronologic data provide support for the interpretation that there was a 600 yr still stand between 14.6 and 14.0 ka.

Two of the samples that were relatively thick tufa showed evidence of depositional hiatuses (Figures A.40-A.41). These horizons were highlighted by brown-colored, irregular surfaces. I interpreted these markers to have formed by subaerial weathering. This implies that the lake was below 2,075 m when they formed and that the lake rose thereafter, during which time the second generation of tufa formed.

Dates on one of the thick samples indicated two separate times of tufa deposition: 14.6-14.1 ka and ~12.2 ka (Figure A.41). A second sample from a different locality showed similar dates: 14.6-14.1 ka; and ~11.8 ka (Figure A.42). Thus the dates imply the lake was below 2,075 m ~14.0-12.2 ka. This inference is supported by lake fluctuation evidence in Section 3 (this appendix).

I have no data on the lake’s elevation between 15.0 ka and 14.6 ka. And because the algal coatings I sampled from thinolite mounds are younger than Ash 1, I am unable to utilize the Wilson Creek tephra to infer lake fluctuation. Thus I cannot determine precisely what the lake level was during this interval. But the U/Th dates on the algal carbonates can still be used to estimate the lake’s elevation when they were deposited. The 15.1 ka algal coatings, therefore, require that the lake regressed from its 2,155-m high stand to 2,075 m during the period encompassing 16-15.1 ka.
Thinolite towers along Cemetery Rd (Figure A.42). I found thinolite mounds at an elevation of 2,015 m to the north of Cemetery Rd. I found white, laminated calcite coatings on the thinolite mounds (Figure A.41). I interpreted the coatings to be algal-mediated deposits. I measured U/Th ages from calcite that coated two separate thinolite mounds. The two analyses yielded dates of 13.6 ka and 13.8 ka. The dates constrain the lake to be 2,015 m at ~13.7 ka. Thus the data imply that the lake fell from 2,075 m to 2,015 m between ~14.1 ka and ~13.7 ka. No lake level indicators are dated for the period encompassing 13.6 and 13.0 kyr.

(Part 3) Mono Lake’s final transgression of the Late Pleistocene (13-12 ka)

My interpretation of the lake’s transgression ~13-12 ka is underpinned by the following: 1) sedimentary data that includes U/Th-dated algal tufa, $^{14}$C-dated charcoal in a sedimentary sequence, and $^{14}$C-dated bird bones found in tufa; and 2) geomorphic data. I reason with this data that 1) Mono Lake transgressed from below 2,010 m to 2,076 m from ~13-12.2 ka, 2) that the lake was at 2,076 m at 12 ka, and 3) that the highest elevation the lake reached during this transgressive period is 2,089 m.

Grand Gash part 2 (Figure A.43). I found two- to three-meter-high thinolite towers rooted at a transgressive disconformity that cross-cuts Ashes 4-1 (Figures A.33, A.34, and A.43). I measured the base of the thinolite mounds to be ~2,015 m. The disconformity can be traced as low as 2,015 m and as high as 2,042 m. I dated two algal coatings on the thinolite towers using
U/Th analysis, which yielded dates of 13.07 ± 0.13 ka and 12.82 ± 0.30 ka. Thus the lake transgressed from below 2,015 m to 2,018 m by ~13 ka.

**Thinolite towers along Highway 167 and Goat Ranch Rd cutoff revisited.** The tufa mounds I previously described in “Thinolite towers along Highway 167 and Goat Ranch Rd cutoff” are revisited here in the context of the lake’s fluctuations 13-12 ka. The tufa mounds there are coated with thick, laminated coatings, which I reason are algal deposits. Some of the coatings show multiple parts. The older parts are dated to 14.6-14.1 ka. I dated the younger part of two separate tufa samples by the U/Th method. The analysis yielded dates of ~12.1 and ~12.2 ka (Figures A.39-A.40). These dates indicate that the lake had risen up to 2,075 m by ~12.2 ka. It, too, was at this elevation at ~12.1 ka.

I also found shorebird bones in a different tufa mound at this locality (Kristie Nelson, personal communication). Because the bones are cemented in the core of the tufa mound, they must be older than the tufa (Figure A.44). Their age can be inferred to indicate when the lake was at 2,075 meters.

I dated two bones using $^{14}$C analysis. This yielded two ages of 10,210 ± 35 $^{14}$C yrs BP (CAMS 170035); and 10,250 ± 35 $^{14}$C yrs BP (CAMS 170101) that calibrate to calendar ages of 11.9 ± 0.08 ka and 12.0 ± 0.09 ka, respectively. The dates indicate that Mono Lake was at 2,075 m ~12 ka. These dates are similar to the U/Th dates on tufa from the same locality. The corroboration between the stratigraphic context and geochronological data strengthen the confidence in my lake fluctuation interpretation.
**Geomorphic evidence for a 2,089-m high stand during the Younger Dryas.** Dates on carbonates and bird bones only constrain the elevation at which the lake must have reached at the time of their deposition. The upper limit of the lake’s rise is less clear. I argue that geomorphic evidence suggests the lake rise terminated at 2,089 m.

When a lake expands and its level rises, an abundance of detritus is produced by wave erosion of the shorelands. If the lake continues to expand, littoral currents will transport this material rangeward (Stine, 1987). And when the lake expansion stops, its level is marked by a berm. The berm will partly comprise the detritus that has been reworked during the lake’s rise. Thus a berm’s size positively correlates with the time of lake rise.

I find no berm of inordinate size between 2,010 and 2,075 m. This suggests that the lake’s rise must have terminated at a higher elevation. Above 2,075 m, there is an anomalously large berm at 2,089 m (Figure A.45). I argue that the 2,089-m berm constrains the termination of the lake’s transgression.

The interpretation that the lake rose to 2,089 m is corroborated by the geomorphology of Black Point. An abrupt cliff marks the northern rim of its welded edifice (Figure A.46). A low-angle surface rises from 2,072 m to its base, which I estimate to be 2,089 m. The low-angle surface and its termination at the 2,089-meter cliff-line suggest that a lake transgression from below 2,072 m terminated at 2,089 m.

The geomorphic and chronologic data support the interpretation that the lake rose from below 2,010 m to 2,089 m. The precise dates on the highest level are unknown. But I can show that the lake rose from 2,010 m to 2,075 m between 13-12.2 ka. And I can show that the lake was at 2,075 meters at ~12.1 ka and ~12 ka.
(4) Fluctuations of the lake during the early Holocene (11-9 ka)

The lake’s fluctuations 11-9 ka are defined by sedimentary sequences exposed along road-cuts as well along the canyon walls of Mill Creek. These data are from two localities: Cemetery Rd; and Lower Mill Creek. $^{14}$C dates from charcoal found in sedimentary sequences and in tufa mounds provide minimum and maximum elevational constraints of the lake, respectively.

My sedimentary observations and interpretations suggest that the lake regressed from some unknown elevation to below 1,985 m before ~11 ka. The lake transgressed above 1,986 m thereafter. And there is evidence to suggest that the lake fluctuated between 1,972 m and 1,978 m during the time interval encompassing 10-6.5 ka. Although there is evidence of lake fluctuation during the early Holocene, their high- and low-stand elevations are unknown. The few data I have collected thus far implies that the lake’s early Holocene fluctuations were within one meter of its late Holocene maximum, 1,981 m; that the lake transgressed above this maximum at least once during the same interval.

Cemetery Rd Locality (Figures A.47-A.48). I found two- to three-meter-high tufa towers approximately 300 m north-northwest of the county cemetery. Physico-chemical tufa compose these towers. This suggests the tufa formed at sublacustrine springs. Their basal elevations range from 1,983 to 1,990 m. I measured the peak of the highest elevation tufa to be ~1,992 m.
I found well sorted lithic pebble gravel and thinolite detritus cemented into one of these towers. I also found charcoal cemented inside of this tufa tower. I reason that the well sorted gravel and thinolite detritus indicate that littoral currents were actively sweeping across this site during the tufa’s formation. Because I did not find an abundance of wood casts in the tufa, which would have suggested vegetation grew proximal to the spring prior to the apparent lake transgression, I infer that the charcoal may have been transported by littoral currents prior to their deposition at the flooded spring site.

I measured the charcoal’s age using $^{14}$C analysis. This yielded a date of $10,960 \pm 490$ a. I interpret the datum along with the sedimentary context to indicate that Mono Lake receded below the base of the tufa mound, $\sim 1,985$ m, and re-rose to an elevation above its crest, $\sim 1,987$ meters, by $\sim 11$ ka. This datum is the oldest evidence of the lake’s early Holocene levels.

I found charcoal in one other tufa mound with a tip of $1,977$ m (Figures A.47-A.48). It is rooted on the upper contact of poorly sorted gravel with detritus up to cobble size. I measured this contact to be $\sim 1,976$ m. The gravel is overlain by 25 cm of carbonate-poor silts with one previously unknown tephra. The sedimentary sequence above the new tephra, which totals approximately 80 cm, comprises silt, but bioturbation thwarts precise observations.

I interpret the gravel as fluvial gravel. And I interpret the silts as lacustrine silts. I thus define this finding upwards sequence to represent a lake transgressive sequence.

I found abundant wood casts and numerous charcoal fragments cemented at the base of the approximately one-meter-high tufa mound. I reason that the best explanation for the coincidence of the wood casts and charcoal at the base of the tufa tower is that the vegetation that once grew proximal to the site of a subaerial spring was later cemented into tufa that formed
following the lake’s inundation of the spring. The age of the charcoal therefore provides an estimate of the time of this lake transgression.

I measured the ages of two separate charcoal fragments from the tufa mound using $^{14}$C analysis. Only one sample was sufficiently large to yield a datum, which was $9,980 \pm 240$ a. This $^{14}$C datum along with its stratigraphic context demonstrates that Mono Lake fell below 1,976 m sometime prior to $\sim 10 \pm 0.4$ ka. The lake then rose above this elevation after this time. It remained above this site until sometime after the deposition of the undescribed tephra.

**Lower Mill Creek Locality (Figure A.49).** I found rhythmic silts that coarsen upwards to massive coarse sands and, ultimately, to massive poorly sorted gravel. The gravel varies in size from pebble to cobble. The contact between the different strata is depositional and not erosional.

I interpret the rhythmic silts as lacustrine silts and the massive coarse sands as sublacustrine debris flow deposits. And I interpret the gravel as fluvial strata. And I interpret this coarsening-upward sequence as a lake regressive sequence.

At the most rangeward exposure of this sedimentary sequence, I identified a change in the bedding orientation of the gravels at 1,972 m (Figure A.49). At this elevation, the bedding geometry changes from massive to moderately dipping to the south. I infer that the transition from the massive gravel to moderately-dipping gravel represents the topset-foreset transition of an ancient delta.

I found three charcoal fragments in the lacustrine silts that just underlie the sandy debris flow deposits. I measured their age by $^{14}$C analysis. The results yielded three dates: $9,830 \pm 90$ a;
9,630 ± 60 a; and 9,510 ± 20 a. I take the youngest age, ~9,500 a, to represent a maximum depositional age for the lake regressive sequence. This age, too, places a maximum age that the lake receded to 1,972 m, which is the lake’s precise elevation that I determined from the deltaic topset-foreset transition.

**The new U/Th-based Mono Lake record compared to previously published curves (Figures A.50-A.51)**

The most recently published Late Pleistocene lake level curve is based on $^{14}$C and paleosalinity proxy data (Benson et al., 1990) (Figure A.50). This record is supported in part by Lajoie (1968)’s interpretations of sedimentary sequences exposed across the Mono Basin. The Benson et al. (1990) record puts forward the following interpretation: that the lake fell from 2,075 m to 2,035 m at 24 ka; that the lake remained at 2,035 m until 17 ka; that the lake rose thereafter to its 2,155-meter high stand, which it reached at ~15.5 ka; that the lake then monotonically fell to 1,965 m between ~15.5 ka and ~11.5 ka, except for a rise from 2,005 m to 2,025 m between 14 and 12 ka; and that the lake level was no more than 10 m above or 20 m below 1,965 m during the remaining part of the Holocene. A comparison of the Benson et al., 1990 lake level interpretation to that from this dissertation shows broad agreement through the record. The records show high stands during the Younger Dryas and Heinrich Stadial 1. But my interpretation shows that the lake began to rise to the Heinrich Stadial 1 high stand several kyr prior to the Benson interpretation. I also show that the lake rose ~60-vertical meters higher during the Younger Dryas than shown for the Benson et al., (1990) interpretation. Where the two records differ greatly is during the Last Glacial Maximum. The Benson et al., (1990) interpretation does not resolve the Big Low. Overall, I would argue that the greatest differences
arise from the greater geologic observations made in this study as well as the significantly higher precision (and, perhaps, accuracy too) of the carbonate U/Th data in this study to Benson et al. (1990)’s lacustrine macrofossil $^{14}$C data.
References


Figure A.1. NASA satellite image showing the Mono Basin. Mono Lake’s perennial tributary streams are highlighted in blue—1) Rush Creek, 2) Lee Vining Creek, 3) Post Office Creek, 4) Mill Creek, 5) Wilson Creek. The orange highlighted areas show the location of the Mono and Inyo Craters. The red-colored box shows the approximate area shown in Figure A.23. The following lettered positions are locations that I sampled material for dating or made stratigraphic observations: (a) type-locality for the Wilson Creek Formation; (b) Grand Gash; (c) “Big Low” and “Big Low” up-cut; (d) Mill Creek; (e) Black Point; (f) Upper Wilson Creek; (g) Highway 167 thinolite mounds; (h) Goat Ranch Rd cut-off tufa mounds; (i) Lower Bridgeport Creek; (j) Cottonwood Canyon; (k) Tufa Crags; (m) carbonate sampling site from the Cowtrack Mountains.
Figure A.2. Photograph looking southwest at the western canyon wall of lower Wilson Creek. The photograph shows fluvial gravels overlain by Lajoie’s Wilson Creek Formation. The colored bars indicate the position of the numbered tephra, 19-1. Ash 1 is eroded from this section. But it is found above Ash 2 further upstream of this position.
Figure A.3. Lajoie (1968)’s Wilson Creek Formation lake hydrograph pinned to the radiocarbon time scale. The lettered intervals refer to the depositional age of sets of Wilson Creek tephra (A: Ashes 1-4; B: Ashes 5-7; C: Ashes 8-15; D: Ashes 16-17; E: Ashes 18-19). The letter “T” refers to the depositional age, $\sim 12 \, ^{14}C \, kyr \, BP$, of the Wilson Creek thinolite that lies above Ash 1. The hydrograph shows that the lake rose from near the present lake level (1945 m) at 30 $^{14}C \, kyr \, BP$. And it then fluctuated between 2,011 m and 2,195 meters (Lajoie marked the overflow elevation at a lower elevation than what it is measured at today). Lajoie inferred that the lowest elevational interval occurred between the deposition of Ashes 5-4--here between Ash Set A (1-4) and B (5-7)--based on the observation of sand beds in this interval at the Wilson Creek Formation type-locality. In addition to this inference, Lajoie surmised that the lake spilled across its southern outlet during the time spanning Ashes 17-19. And he reasoned that the 2,155-meter high stand of the lake occurred during the time after the deposition of Ash 1. And he suggested that the lake fell abruptly after this high stand.
Figure A.4. (a) Photograph looking east towards Black Point showing tufa mounds above the shoreline. (b) Photograph looking northwest from the lake’s southwestern shore. It shows tufa mounds rooted at or below the present lake level.
Figure A.5. (a) Photograph showing thinolite crystals radiating outwards from a nucleus, which is the orifice of an ancient spring. (b) a photograph showing a closer view of the thinolite crystals.
Figure A.6. Photograph showing a sedimentary sequence exposed along the eastern canyon walls of Upper Wilson Creek. The sequence comprises Wilson Creek silts intercalated with Ash 1. A thinolite mound overlies Ash 1. It is common to find this sequence throughout the Mono Basin.
Figure A.7. (a) Photograph showing a bipartite tufa mound from a time within the interval encompassing the deposition of Ashes 15-19. Its inner part comprises a porous, physico-chemical core. And its outer part comprises dense, laminated calcite, which I interpret to have formed by algae. (b) Photograph showing a modern example of the tufa pictured in (a). I presumed the dark-green-colored laminae are algae. The tan-colored material comprises laminated tufa. I am unsure what the translucent prismatic crystals are.
Figure A.8. (a, b) Photograph showing algal tufa that grew first as planar parallel laminations and then in the form of upward-growing domes.
Figure A.9. Photograph showing “dirty” stromatolites over- and under-lain by silts.
Figure A.10. Photograph looking northwest from the Mono Craters towards the Sierra Nevada. The morainal complexes of the six glacial streams that flowed into the Mono Basin are highlighted by the red-dashed line. These glacial systems are Grant Lake (GL), Parker Creek (PC), Sawmill Creek (SC), Bloody Canyon (BC), Lee Vining Canyon (LVC), and Lundy Canyon (LC).

Figure A.11. Photograph looking southeast at the eastern canyon wall of Mill Creek. The red-circled areas show the approximate location of two sites that I made sedimentary and stratigraphic observations: (1) Big Low type-locality; (2) Big Low up-cut
**Figure A.12.** Photograph showing the eastern canyon walls of Mill Creek of our “Big Low” type-locality. Labels S1, S2, S3 refer to units that comprise lacustrine silt. Labels G1 and G2 refer to units that comprise gravels (fluvial and lacustrine). And the dashed lines show the stratigraphic position that we conclude must represent disconformities. The lettered, white boxes show the approximate position of photographs that are examined further herein.

**Figure A.13.** (box "a" in A.12) Photograph showing S1 overlain by S3. A disconformity, which is shown by the white line, marks the two layers. Here S3 comprises cobble gravel, silt laminae, and massive to coarsening upward very fine to fine sands. The scale in the photo is a 28-cm-long ruler.
Figure A.14. (box “b” in A.12) Photograph showing S1, G1-1, G1-2, G2, and S2. The contacts shown by the solid and dashed lines are disconformities. Ash 14 is partly shown in this photograph, but its lower contact is not clear. Ash 15 is also found in S1, but it lies below the observable field.
Figure A.15. (box “c” in A.12) Photograph showing G1-1, G1-2, S2. Ash 2 is clearly observed. Ash 4 and 3 are not perceivable in this photo, but they are indeed found below Ash 2 (see box “d”). A grey- and blue-colored sharpie pen is in the G1-1 area for scale.
Figure A.16. (box “d” in A.12) Photograph showing Ash 4, 3, and 2 in lacustrine silts(S3). The scale in the photo is a yellow-colored meter stick.
Figure A.17. (box “e” in A.12) Photograph showing G1-1, G1-2, and the disconformity (the white line) that I mark between the two.
Figure A.18. (box “f” in A.12) Photograph showing the gravel units that compose G1--1, 2, 3, and 4—and S3, the lacustrine silt that contains Ashes 4, 3, 2, and 1. The white lines show the contacts between the unit. I interpret each contact to be a disconformity except for the contact between G1-3 and G1-4 and between G1-3 and G1-2. I infer this contact is depositional. The red box labeled “g” shows the position of box “g” in Appendix Figure12. I use a white-colored, 28-cm-long rule, which is in the area highlighted by the red box, for scale.
Figure A.19. (box “g” in A.12) Photograph showing a closer view of G1-2, G1-3, and G1-4. G1-2 and G1-4 are fluvial cobble gravel. G1-3 is a littoral embankment. The contacts between the G1 units, which are depositional, are marked by white lines. The dashed white line between G1-3 and G1-4 shows the approximate location of where the contact should be. The G1-3 deposit obscures its actual position. The results of the U/Th analyses measured on calcite cementing the G1-3 conglomerate (S49) is shown on the figure.
Figure A.20. (photograph “h” in A.12) Photograph showing the silt block, S2, that was deposited within G1-1 and overlain by G1-2. S2 is strongly convoluted. I found Ash 7, Ash 6, and Ash 5 in S2; however, Ash 7 is the only clearly recognizable tephra in this photograph. Ash 6 and Ash 5, which measure only a few-millimeters-thick cannot be observed at this distance.
Figure A.21. (photograph “I” in A.12) Photograph showing Ash 5 and Ash 6 in S2. I reason the double layer of Ash 6 is a product of prior deformation that resulted in convoluted stratigraphy. This is consistent with observations I made of the Ash 7, 6, and 5 sequence in sedimentary exposures that are no more than 100 meters up- and down-stream of this location.
Figure A.22. Photograph showing the wave-cut disconformity (red-dashed line) cross-cutting lacustrine silts that contain Ashes 5-7. And it shows the littoral sands that overlie this disconformity. Lacustrine silts overlie the littoral sands.
Figure A.23. Photograph showing Mill and Wilson Creek and the northern half of Black Point. Location #1 is the “Big Low” type-locality, which is show in Figures A.11-A.12. Location #2 is the location of the observed sedimentary sections from Lower Wilson Creek. Location #3 shows the position of the observed sedimentary section from Upper Wilson Creek.
Figure A.24. Photograph showing one variant of the Big Low sands (Sd1) that I find overlying lacustrine silt containing Ashes 19-5 (S1) and Ashes 4-2 (S2). The sand bed in this photograph is approximately two-centimeters-thick.
Figure A.25. Photograph showing the western canyon wall of Lower Wilson Creek and the upper half of Lajoie’s Wilson Creek Formation type-locality. Two lacustrine silt units, S1 and S2, are bisected by Sd1, a well-sorted fine to medium sand layer. I found Ash 5, 6, and 7 in S1. And I found Ash 2, 3, and 4 in S2. Ash 1 is not found at this location because of a younger, Holocene lake rise that beveled it along with a portion of Ash 2. Ash 6 is too thin to be observed in this photo. I reason that Sd1 represents littoral deposition of sands during the lake’s transgression following the Big Low. This, then, requires that the contact between Sd1 and S1 to be a disconformity (indicated here by the red-dashed line)—more specifically, an erosional surface as a result of wave planation. The contact between Sd1 and S2, which is broadly a fining upwards sequence, is depositional. The crude north-dipping orientation of the lower half of S1 shows the deformed section of S2.
Figure A.26. Photograph showing the Big Low sands (Sd1) between lacustrine silts containing Ashes 19-5(S1) and Ashes 4-1(S2). Sd1 is thicker at this location than in the prior photograph. It is also more strongly deformed.
Figure A.27. Photographs “a” and “b” show the location (red circle) on the tufa mound that I sampled the calcite from in the field. Photograph “c” shows the crystals (UWC5-1A and UWC5-1B) that were analyzed using the U/Th method. The resultant data is shown, too.

UWC5-1A: 20.46 ± 0.24 ka
UWC5-1B: 20.21 ± 0.19 ka
Figure A.28. Photograph showing the 20.5 ka tufa mound’s stratigraphic relationship. Dr. Xianfeng Wang is pointing here to where we found Ash 4 overlying the sampled tufa mound.
Figure A.29. Photograph showing Ashes 4-2 in lacustrine silts that abut the 20.5 ka tufa mound.
Figure A.30. Photograph showing the north gully wall of Lower Bridgeport Creek. We grouped the lacustrine silt strata here into three units: S1, which contains Ashes 8-5; S2; and S3, which contains Ashes 4-2. Ash 3 and 2 are not seen in this photograph. S2 is unique. It is exceptionally rich in carbonate. And it contains calcareous laminae, which manifest as protuberant ridges. Because I infer that the expression of the “Big Low” must be between Ash 5 and 4, I attribute my observations of S2 to the low lake conditions. I have not determined a reasonable explanation for S2’s character. But I find no evidence of subaerial exposure in this sequence, which is measured at 1,955 meters.
Figure A.31. Photograph looking north towards the Tufa Crags. The buff-white-colored outcrops are lacustrine silts that contain Wilson Creek Formation tephra.
Figure A.32. Photograph showing the dated calcite coatings that were sampled from the Tufa Crags. The brown-colored material is the physio-chemical tufa that composes the structure of the Tufa Crags. The white-colored material coating the physico-chemical is what I dated using the U/Th method. Note the speleothem-like structure of the coating adjacent to S549.

Figure A.33. Photograph looking SE at the eastern wall of Grand Gash. The faint white-grey surface in the right-middle of the photograph is Mono Lake. The locations of Sequence A and Sequence B are indicated by the white-arrowed line. The photograph shows the fluvial cobble gravel (G1) overlain by the lacustrine silts (S1). The yellow line indicates the contact between
the two units. S1 is not found in Sequence A, which is precisely where the ~13-kyr thinolite outcrops at the surface. The U/Th ages from calcite coating the thinolite towers is shown by the pink-colored highlights. The transgressive disconformity we identified is indicated by the solid, pink line.

Figure A.34. Photograph looking East at the eastern wall of Grand Gash. The locations of Sequence A, B, and C are indicated by the white-arrowed lines. The photograph shows the fluvial cobble gravel (G1, blue) overlain by thinolite (pink) in Sequence A and by lacustrine silts (S1, green) in Sequence B. Sequence C, which comprises riverine gravels and sands (G2, orange), abuts G1 and S1. The respective U/Th ages from calcite in G2 and the thinolite that overlies G1 is indicated by the orange- and pink-colored highlights, respectively. The transgressive disconformity I identified is indicated by the solid, pink line.

Figure A.35. Photographing looking east at the top of the G1 gravels in Sequence C. The south-dipping bedding is apparent in the conglomerate. The U/Th-dated calcite was sampled from the conglomerate exposed at the top of this outcrop.
Figure A.36. Photographs showing the location of calcite sample S559 (photo on right) and isopachous character of the sample (photo on left)
Figure A.37. (a) Photograph showing the sampled calcite. (b) Photograph showing the stratigraphic context of sample SN-3-2.

SN-3-2 (1): $15.94 \pm 0.05$ ka
SN-3-2 (2): $15.90 \pm 0.06$ ka
Figure A.38. (a) Photograph showing fractures on Black Point with its walls coated by dense, white, laminated calcite. (b) Photograph showing sampled calcite that I U/Th dated. (c) Photograph showing the analytical results—$17.29 \pm 0.20$ ka—of the dated sample.
Figure A.39. (top-left photo) Photograph looking at the eastern escarpment of the Cowtrack Mountains at the sampling locality. (top-right photo) Photograph of the isopachous calcite cement that I sampled for U/Th analysis. (bottom-left and bottom-right photos) Photographs showing the two samples and their respective dates that were measured by the U/Th method.
Figure A.40. Photograph of algal tufa I collected at an elevation of 2,075 m. It coated a physico-chemical tufa mound. The algal tufa sample shows three separate intervals of calcite deposition separated by hiatus marked by irregular surfaces, one of which is brown-colored and the other grey. But the U/Th dates show two distinct periods of calcite deposition: 14.4-14.3 ka; and 12.2 ka.
Figure A.41. Photograph showing an algal tufa sample I dated using the U/Th method. This sample shows two intervals of deposition: one ~14.5-14.2 ka; and a younger period ~11.8 ka. There is a grey-colored zone between the two distinct age groups, which I infer to be a result of subaerial weathering. The groups of U/Th ages are nearly identical to the sample in Figure A.40.
Figure A.42. (a) Photograph looking southeast from the Cemetery Rd thinolite mounds. Upper-left and bottom-left photos show the dense, white, laminated calcite samples--CR-1A and CR-1B--that coats the physico-chemical tufa mound.

CR-1A: 13.64 ± 0.14 ka
CR-1B: 13.83 ± 0.23 ka
Figure A.43. Photograph looking north and up Grand Gash. It shows an example of a thinolite mound underlain by cobble conglomerate and cobble gravel. I dated algal calcite coatings on the thinolite to ~12.8 and ~13.1 ka.
Figure A.44. (a) Photograph showing the bird bones (indicated by the red star) embedded in the physico-chemical matrix of a tufa mound. (b) Photograph showing a closer view of the bones cemented into the tufa after I removed it from the mound’s structure. Two bones were dated by $^{14}\text{C}$ analysis. These analyses yielded dates of $11,920 \pm 80$ a and $12,000 \pm 90$ a (CAMS 170035 and CAMS 170101, respectively).
Figure A.45. Aerial photograph showing a succession of littoral embankments that are prominent features across Mono Valley, the low-relief lake plain of the eastern Mono Basin. The lowest elevation berm in this image is 2,048 m. The highest is 2,150 m. They represent still-stands of the lake that track its recession from the 2,155-meter high stand.

Figure A.46. Photograph looking northeast from Warren Bench towards Black Point. The white-dashed line is the most prominent wave-cut cliff-line on Black Point. It is a Holocene feature. The yellow-dashed line marks the approximate location of a similar cliff-line at ~2,089 m.
Figure A.47. Photograph looking northeast at the southern exposure of a one-and-half-meter-thick sedimentary sequence exposed along a road cut. I collected charcoal from the base of the tufa mound shown in this photograph.
Figure A.48. Photograph of the tufa fragment that I sampled from the mound in Figure A.47. I
$^{14}$C dated the charcoal I found in this tufa fragment. Its date is $9,980 \pm 240$ (CAMS 171893).
Figure A.49. Photograph looking northeast at the sedimentary sequence exposed along the eastern canyon walls of Mill Creek. The dashed line shows the change in bedding of the cobble gravel—from horizontal to moderately dipping to the south. They represent the topset and foreset beds of the ancient Mill Creek. And the transition between the topset and foreset beds marks the precise level of the former lake.
Figure A.50. Benson et al., (1998)’s interpretation of Mono Lake level record using data from Benson et al., (1990). The black line replicated the study’s interpretation.
Figure A.51. My lake level record (upper plot) compared to Benson et al. 1998’s interpretation (lower plot). The blue-highlighted intervals encompass the time of inordinate cooling in the North Atlantic—Heinrich Stadial 2 (HS-2, 24.5-23.5 ka), Heinrich Stadial 1 (HS-1, 18.6-14.6 ka), Younger Dryas (YD, 12.9-11.5 ka). In a broad sense, the two records both show that the lake rose during Henrich Stadial 1 and Younger Dryas. Both show that the lake regressed from its 2,155-meter high stand to a similar low stand of ~2,010 m during the Bolling-Allerod (BA, 14.6-12.9 ka), a period of anomalous warmth in the North Atlantic. And both show that the lake was lower than its deglacial elevational range after the Younger Dryas. But the two records show some significant differences—namely, that the Benson et al., 1990 record does not resolve the Big Low and underestimates the high stand elevation during the Younger Dryas.