



# The Late Holocene Dry Period: multiproxy evidence for an extended drought between 2800 and 1850 cal yr BP across the central Great Basin, USA



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## ABSTRACT

Evidence of a multi-centennial scale dry period between ~2800 and 1850 cal yr BP is documented by pollen, mollusks, diatoms, and sediment in spring sediments from Stonehouse Meadow in Spring Valley, eastern central Nevada, U.S. We refer to this period as the Late Holocene Dry Period. Based on sediment recovered, Stonehouse Meadow was either absent or severely restricted in size at ~8000 cal yr BP. Beginning ~7500 cal yr BP, the meadow became established and persisted to ~3000 cal yr BP when it began to dry. Comparison of the timing of this late Holocene drought record to multiple records extending from the eastern Sierra Nevada across the central Great Basin to the Great Salt Lake support the interpretation that this dry period was regional. The beginning and ending dates vary among sites, but all sites record multiple centuries of dry climate between 2500 and 1900 cal yr BP. This duration makes it the longest persistent dry period within the late Holocene. In contrast, sites in the northern Great Basin record either no clear evidence of drought, or have wetter than average climate during this period, suggesting that the northern boundary between wet and dry climates may have been between about 40° and 42° N latitude. This dry in the southwest and wet in the northwest precipitation pattern across the Great Basin is supported by large-scale spatial climate pattern hypotheses involving ENSO, PDO, AMO, and the position of the Aleutian Low and North Pacific High, particularly during winter.

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## 1. Introduction

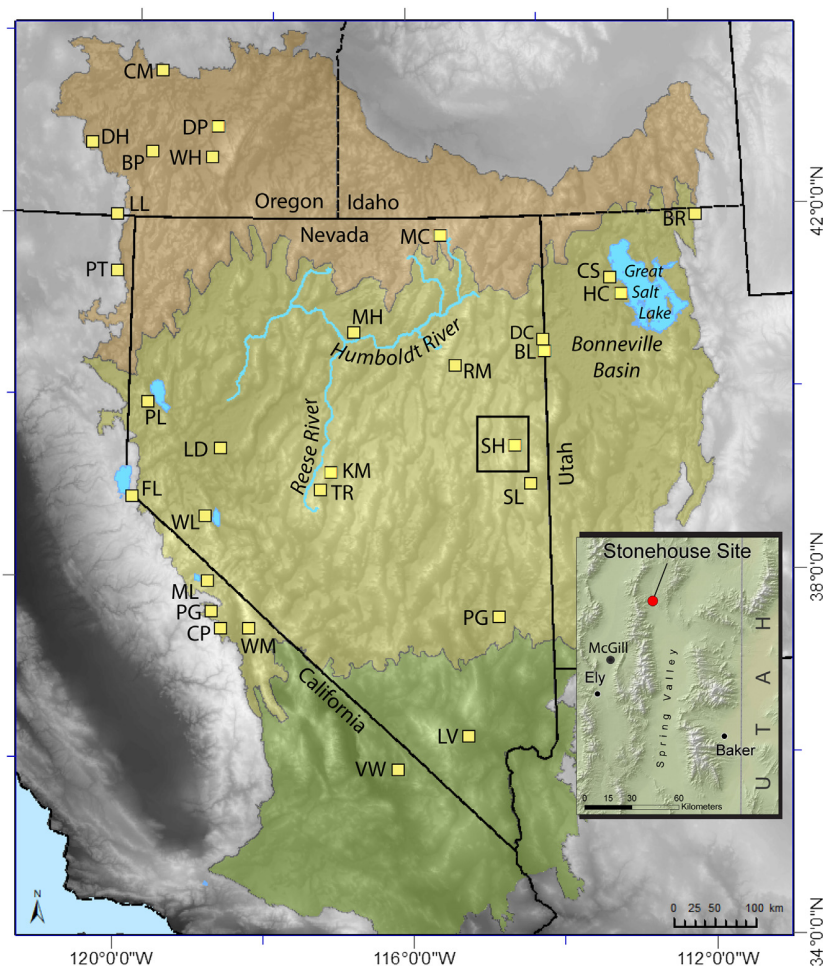
Paleoclimate reconstructions of centennial to multi-centennial long dry periods are important indicators of persistent Holocene ocean/atmosphere circulation patterns. Two periods of severe and persistent Holocene drought are well-documented in the Great Basin of the western United States: the middle Holocene dry period about 7500–5000 cal yr BP (Lindström, 1990; Grayson, 2000; Benson et al., 2002; Grayson, 2011) and the Medieval Climate Anomaly between ~1200 and 750 cal yr BP (Stine, 1994; Kleppe et al., 2011). Evidence for Middle Holocene aridity comes from a number of independent proxies across the Great Basin region, including lowered lake level and an increase in drought tolerant vegetation at Pyramid Lake, Nevada (Benson et al., 2002; Mensing

et al., 2004), submerged tree stumps in Lake Tahoe, California (Lindström, 1990), lowered water tables and reduction of marshland in the Ruby Marshes, eastern Nevada (Thompson, 1992), and Diamond Pond, southeastern Oregon (Wigand, 1987), decline in small mammal diversity at Homestead cave in north-central Utah (Grayson, 2000) and changes in treeline in the White Mountains, eastern California (LaMarche, 1973). Middle Holocene aridity in western North America has been attributed to expansion of the east Pacific subtropical high-pressure system and increased summer radiation in the northern hemisphere associated with variations in Earth's tilt and the timing of perihelion (Whitlock et al., 1995; Whitlock and Bartlein, 1997; Milsaugh et al., 2000).

Severe and persistent drought is similarly well documented for the Medieval Climate Anomaly, including submerged tree stumps in Mono Lake, eastern California (Harding, 1965; Stine, 1990, 1994) the Walker River, Tenaya Lake, Osgood Swamp, Nevada and California (Stine, 1994), and Fallen Leaf Lake, California (Kleppe et al., 2011), lowered lake level and an increase in drought tolerant

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**Fig. 1.** Map of the Great Basin ecoregion, divided into the northern (brown), central (olive) and southern (green) ecotypes. Black rectangle defines the study area and inset map provides a large scale map of Stonehouse Meadow. Sites referred to in the text are marked with yellow squares and acronyms as follows: BL (Blue Lake), BR (Bear Lake), CM (Craddock Meadow), CP (Cirque Peak), DC (Danger Cave), DH (Dead Horse Lake), DP (Diamond Pond), FL (Fallen Leaf Lake), HC (Homestead Cave), KM (Kingston Meadow), LD (Lead Lake), LL (Lily Lake), LV (Las Vegas Valley), MC (Mission Cross Bog), MH (Middle Humboldt River), ML (Mono Lake), PG (Palisade Glacier), PG (Lower Pahranaagat Lake), PL (Pyramid Lake), PT (Patterson Lake), RM (Ruby Marsh), SH (Stonehouse Meadow), SL (Stella Lake), TR (Toiyabe Range), VW (Valley Wells), WH (Wildhorse Lake), WL (Walker Lake), WM (White Mountains).

vegetation at Pyramid Lake (Benson et al., 2002; Mensing et al., 2004), lowered lake levels in Walker Lake (Yuan et al., 2004, 2006; Adams, 2007), increased temperature and decreased precipitation reconstructed from tree-rings along the eastern Sierra Nevada (Graumlich, 1993), and reduced stream-flow in the Sacramento River, California (Meko et al., 2001). The cause for extended drought during this period is not well understood, although a persistent ridge of high pressure over the western United States has been proposed (Stine, 1994; Graham, 2004).

Much less attention has been paid to another potentially persistent Late Holocene Dry Period between ~2800 and 1850 cal yr BP. Tausch et al. (2004) described regional evidence for what they termed the post-Neoglacial drought between 2600 and 1600 cal yr BP. Mensing et al. (2007) argued for a dry period between 2500 and 1800 cal yr BP and noted that evidence appeared stronger in the western Great Basin than in the eastern portion. In a recent review of Great Basin climate history, Grayson (2011) discussed a potential dry period between 2500 and 1900 years ago, but indicated that the scope and magnitude have yet to be determined.

The purpose of this paper is to refine our understanding of the temporal and geographic extent of this drought. We present new

pollen, diatom and mollusk data from Stonehouse Meadow in Spring Valley, eastern Nevada that provides evidence of a multi-centennial scale dry period between ~2800 and 1850 cal yr BP, and compare these results with other Great Basin studies. Multi-centennial dry periods provide an opportunity to describe the effects of long-term climate patterns on hydrologic and biotic systems. Extended droughts are also of concern to water managers throughout the western United States tasked with providing regular water supplies to urban areas. Regional aridity and the paucity of surface water in many states in the western U.S. make municipalities and agriculture increasingly reliant on groundwater. For example, the city of Las Vegas purchased groundwater rights in Spring Valley (~250 miles north of the city) for potential export out of the basin. It is critical to have an accurate reconstruction of the magnitude and duration of past droughts and their effect on the landscape to better understand and plan for future droughts.

## 2. Study area

The hydrogeology of Spring Valley (Fig. 1) is dominated by both valley-fill and carbonate-rock aquifers. Valley-fill aquifers are unconsolidated deposits composed of boulders, gravel, sand, silt, and

clay that have been eroded from the adjacent mountain blocks and are thousands of feet thick near the center of the valley (Mankinen et al., 2006). These valley-fill deposits form high permeability aquifers in the valleys of eastern Nevada and western Utah (Plume, 1996; Welch et al., 2007). Carbonate-rock aquifers underlie most of eastern Nevada, including the Spring Valley area, and often form deep regional flow systems that encompass several valleys (Thomas et al., 1986; Plume, 1996; Hershey et al., 2007; Welch et al., 2007). The carbonate rock aquifers are thousands of feet thick and provide a connection for groundwater flow from mountain block recharge areas to valleys (Plume, 1996; Thomas et al., 1996; Mankinen et al., 2006; Lundmark et al., 2007; Welch et al., 2007).

Melting winter snowpack supplies the majority of groundwater recharge by infiltrating fractures in the consolidated rock in the mountains as well as from infiltrating stream flow on alluvial fans (Welch et al., 2007). Tritium/helium data show that it takes one-to-four years for groundwater recharge to reach valley springs (Hershey et al., 2007). Gates (2007) found a direct relationship between precipitation amount and groundwater levels in the adjacent Snake Valley. During five regional droughts (1930–1936, 1953–1965, 1974–1978, 1988–1993, and 1999–2004) less-than-average precipitation lowered groundwater levels, whereas the wet period of 1982–86 resulted in greater-than-average recharge and raised the groundwater level.

Modern precipitation is characterized by a spring maximum (31% of annual precipitation) and the remainder falls fairly equally between the other three seasons: 21% in winter; 25% in summer; 23% in fall (Benson and Klieforth, 1989). Average January and July maximum temperatures are 3.9 °C and 30.5 °C, respectively with annual precipitation of 224.8 mm at McGill (1935 m elev.), the closest climate station to the study area (WRCC, 2013). During winter months, an increasingly larger fraction of the annual total precipitation occurs at higher elevations (Thomas et al., 2012).

Stonehouse Meadow in Spring Valley (39° 47' 8"N, 114° 32' 34"W) is a spring-fed sedge (*Carex* spp) meadow (~1740 m elev.). The spring supports populations of the aquatic mollusk *Pyrgulopsis kolobensis* as well as other terrestrial and aquatic species. *P. kolobensis* is an obligate spring-dwelling (crenobiontic) species although it can survive being out of water for a few hours (McCabe, 1998; Sada and Herbst, 2006).

The vast majority of Spring Valley is dominated by native vegetation communities representative of the Great Basin (Lowry et al., 2005), including *Artemisia tridentata* (big sagebrush), *Atriplex* spp. (salt desert scrub) and scattered *Ephedra* sp. (Mormon tea). The meadow is grazed by cattle during summer.

*Pinus monophylla* (pinyon pine) and *Juniperus osteosperma* (Utah juniper) are the dominant trees in the nearby mountains with *Cercocarpus* sp. (mountain mahogany) also present. The pinyon–juniper woodland lower elevational limit averages about 2400 m. Subalpine conifers present in the adjacent Schell Creek Range include *Abies concolor* (white fir) in pockets above 2600 m, and scattered *Pinus flexilis* (limber pine), *Pinus longaeva* (bristlecone pine) and *Picea engelmannii* (Engelmann spruce) above 3000 m elevation (Charlet, 1996).

### 3. Methods

#### 3.1. Sediment coring

Duplicate adjacent overlapping 7.0 m long sediment cores (STH-10-2 and STH-10-3) were recovered in 2010 with a modified Livingstone corer. The surface sediments were composed of rooted plants and undecomposed roots that make a nearly impenetrable fibrous mat. The upper 30 cm of this material was removed with a hand saw and bagged before coring. Some compaction of sediments

occurred during coring. The difference between the distance pushed and length of sediments recovered was measured and actual sediment depth was calculated by comparing the total recovery with the amount compacted assuming constant compaction. Between 363 and 380 cm depth, 17 cm of core was lost from STH-10-3. Cores were extruded into plastic tubes, and transported to the University of Nevada, Reno (UNR).

#### 3.2. Sediment analyses

All cores were described and photographed, then stored at 4 °C in the UNR Geography Palynology Laboratory. Twelve samples (bulk sediment, mollusk, and seeds) were selected for radiocarbon dating. Percent total organic matter (%TOM) was measured using loss on ignition analysis from samples taken from contiguous 1-cm samples (1.25 cc) for the full length of STH-10-3. Samples were oven dried at 100 °C for a minimum of 24 h, weighed and then placed in a furnace at 550 °C for one hour to combust all organic matter (Dean, 1974). Samples were cooled, weighed and then placed in a 1000 °C furnace for one hour to obtain percent total carbonate (%TC). Sediment type (clay, silt and sand) was qualitatively described by feel.

##### 3.2.1. Pollen analyses

Twenty-six samples were removed for pollen analysis (0.625 cc). Samples were taken at ~25 cm intervals, and more intensive sampling was done where greater detail was desired. Pollen preparation used standard chemical digestion methods following Faegri and Iversen (1985). A known quantity of exotic tracers (*Lycopodium* spores) was added to each sample for calculating absolute pollen concentration (Stockmarr, 1971). Samples were sieved with 180  $\mu$  mesh following potassium hydroxide (KOH) digestion to remove large organic fragments (generally peat and roots).

Pollen was counted using 400 $\times$  magnification. The goal was to count a minimum of 400 terrestrial pollen grains; however for strata with very low accumulation rates this was impractical. In such cases ( $n = 13$ ), we counted a minimum of 200 grains. Pollen was counted by Mensing, Tunno and Smith. The investigators recounted select samples of each other's work to verify inter-operator reliability, and were able to reproduce the same results. Pollen was identified to the lowest possible taxonomic level using reference material in the UNR Geography Palynology Lab and published pollen keys (Moore and Webb, 1978; Kapp et al., 2000). Pollen of the families Cupressaceae and Taxaceae was assumed to be *Juniperus* (juniper), since no other genus of these families occurs in the region.

Pollen percentages were calculated from the sum of terrestrial pollen, excluding the wet meadow taxa Cyperaceae and Potamogeton. Accumulation rates (grains cm<sup>-2</sup> yr<sup>-1</sup>) were calculated by dividing concentration (grains cm<sup>-3</sup>) by the number of years per sample (yr cm<sup>-1</sup>). Cyperaceae represented nearly 100% of the wet meadow pollen, therefore we plotted this taxon by accumulation rate rather than percentage so that changes in abundance would be apparent. Zonation was interpreted from a constrained single-link dendrogram created using a cluster analysis program modified from Birks and Gordon (1985); data input included the twelve most common taxa plus %TOM and %TC.

##### 3.2.2. Mollusk analyses

For mollusks, 5 ml was removed from core STH-10-3 within a 1 cm-thick slice of the core every 5 cm. Samples were sieved with deionized water through nested screens (smallest 250 microns) and air dried. Samples containing primarily clay (666–620 cm, 540–510 cm, 255–235 cm) were placed in 600 ml deionized water with 1.25 ml sodium hexametaphosphate and soaked for

approximately 30 min to disaggregate clays. Sediments were rinsed in deionized water until a neutral pH was achieved. Mollusks were picked from the sediment matrix using a Nikon microscope at 10× or greater and identified. Identifications were made using the author's (Sharpe) reference collection, Bequaert and Miller (1973), Burch (1975), Clarke (1981), and Burch (1989). Some identifications were verified using the mollusk reference collection at the Monte L. Bean Life Science Museum, Brigham Young University. Taxonomy used here follows USDA Integrated Taxonomic Information System (<http://www.itis.gov>).

### 3.2.3. Diatom analyses

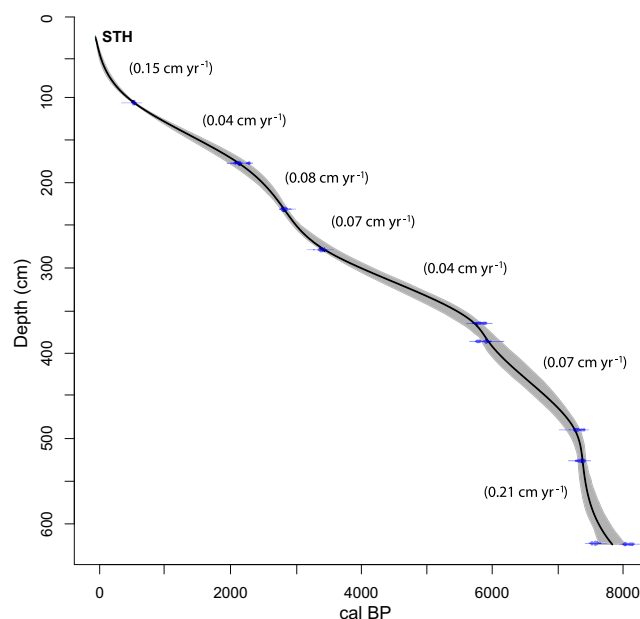
Samples for diatom analysis were processed from twenty-seven depths between 180 and 600 cm using hydrogen peroxide, hydrochloric acid, and nitric acid (Battarbee, 1986) to remove carbonate and organic matter, and sodium pyrophosphate to deflocculate clay. Approximately 50 µl of the resulting suspension was dried on a 22 × 30 mm coverslip and permanently mounted using Naphrax. Test counts indicated that the diversity of the diatom flora was relatively low. Therefore, 300 frustules were counted following the method of Schrader and Gersonde (1978) using a microscope with Nomarski optics at magnifications of 630× and 1000×. Diatom identification was based on Lowe (1974), Gasse (1986), Krammer and Lange-Bertalot (1986, 1988, 1991a, 1991b), Lange-Bertalot and Krammer (1987, 1989), Cumming et al. (1995), Round and Bukhtiyarova (1996), Krammer (1997a, 1997b, 2000, 2002) and Lange-Bertalot (2001).

## 4. Results and interpretation

### 4.1. Chronology

Twelve AMS radiocarbon dates were obtained on plant macrofossils, terrestrial mollusks (*Vallonia*), and bulk sediments. All  $^{14}\text{C}$  dates (Table 1) were calibrated using Calib 6.0 using the INTCAL 2009 dataset (Reimer et al., 2009). An age of 'modern' was obtained on bulk organic sediment from the surface, and we assumed from this that there were no carbon reservoir effects down core. This was further tested with a set of duplicate dates, one on bulk sediment and one on a plant macrofossil at 389 cm depth. These produced statistically identical ages. The two basal dates did return different ages, and for the age chronology we took the average of the two, placing the base of the core section analyzed (660 cm) at 8000 cal yr BP.

Our age model (Fig. 2) was constructed using a smoothed spline in the Clam 2.1 program (Blaauw, 2010). The 'best' age assigned by the program for each centimeter depth was used for plotting data,



**Fig. 2.** Age model for Stonehouse Meadow. Black line represents the best fit used for plotting data. Blue bars represent calibrated radiocarbon ages with the two sigma error. Gray band represents the 95% confidence interval calculated using clam 2.1 (Blaauw, 2010). Average sedimentation rate between radiocarbon ages is shown except for short intervals with overlapping dates. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

and the 95% confidence interval was used to determine potential age range. Age dates referenced from other studies were calibrated using the median probability for the two sigma range calculated in Calib 6.0 (Reimer et al., 2009). All ages are presented as cal yr BP. Sediment accumulation rates are fairly consistent through most of the core, averaging between 0.04 and 0.08 cm yr<sup>-1</sup>, but are higher near the bottom and top of the core.

### 4.2. Sediments and zonation

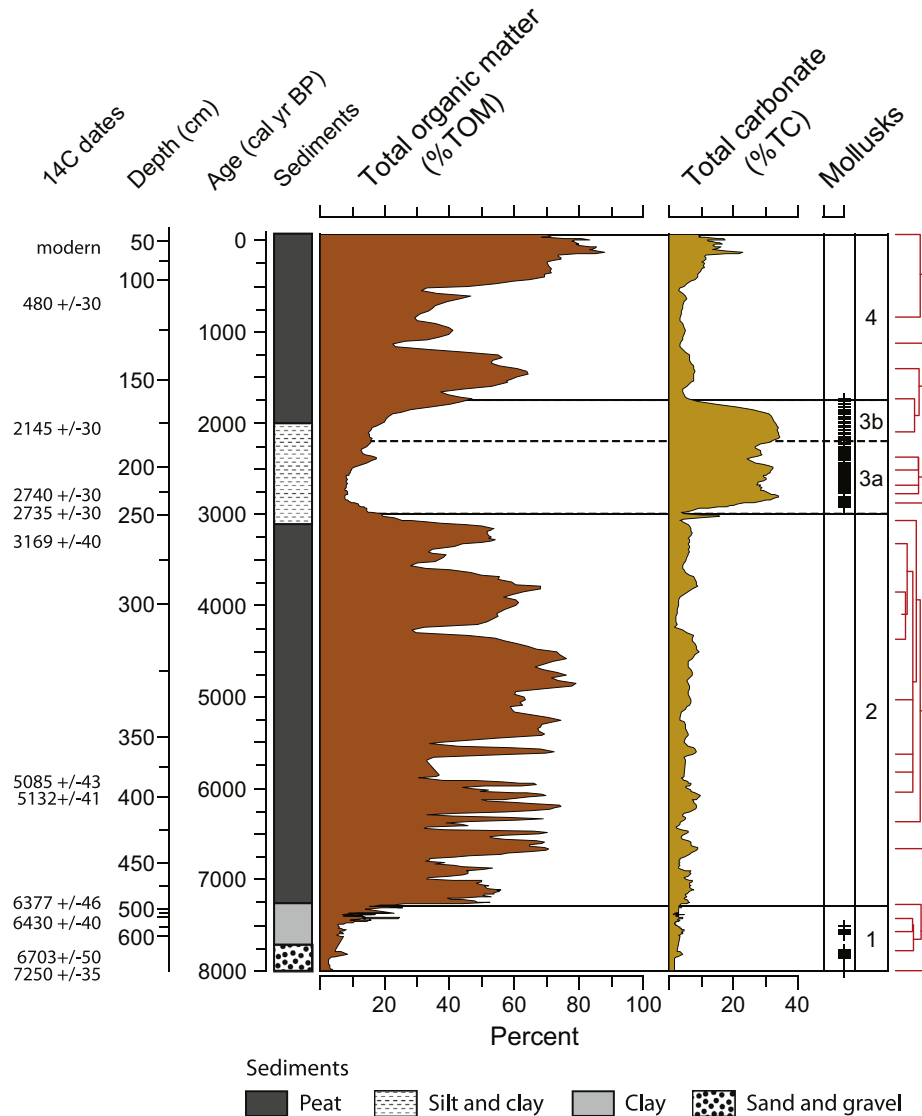
The upper 50 cm consisted of a dense root mass. From 50 to 100 cm, sediments are composed of dark organic rich peats (Fig. 3) with an average sedimentation rate of 0.15 cm yr<sup>-1</sup> (~7 yr cm<sup>-1</sup>). Peats continue to 175 cm depth, although the sedimentation rate slows to 0.04 cm yr<sup>-1</sup> (~23 yr cm<sup>-1</sup>). From 175 cm to 257 cm depth, sediments become rich in silts and clays and the sedimentation rate increases to 0.08 cm yr<sup>-1</sup> (~12 yr cm<sup>-1</sup>). Below

**Table 1**  
Stonehouse Meadow radiocarbon dates. Shaded cells represent paired discrete/bulk samples.

Stonehouse	Sample depth	Lab number <sup>a</sup>	$^{14}\text{C}$ Age	Cal yr B.P.			$\delta^{13}\text{C}$	Material
				yr ±	2Σ min	Median prob.	2Σ max	
STH-10-3	66–67 cm	AA91554	Modern			Modern	–28.7	Bulk organic
STH-10-3	115–116 cm	Beta 300431	480	30	500	520	–25.5	Sedge seeds
STH-10-3	185–186 cm	CAMS152711	2145	30	2010	2140	–	Mollusk: <i>Vallonia</i>
STH-10-3	235–236 cm	CAMS152712	2740	30	2770	2830	–	Mollusk: <i>Vallonia</i>
STH-10-3	245–246 cm	CAMS152713	2735	30	2760	2820	–	Mollusk: <i>Vallonia</i>
STH-10-3	261–262 cm	AA91555	3169	40	3270	3400	–26.9	Bulk organic
STH-10-3	388.5–391 cm	AA91556	5085	43	5730	5820	–28.2	Plant
STH-10-3	389–390 cm	AA91557	5132	41	5750	5880	–27.2	Bulk sediment
STH-10-3	490–491 cm	AA91558	6377	46	7120	7310	–25.9	Bulk sediment
STH-10-3	530–531 cm	Beta 300434	6430	40	7280	7360	–25.2	Sedge seeds
STH-10-3	626–627 cm	AA91559	6703	50	7490	7570	–	Bulk sediment
STH-10-3	626–627 cm	CAMS152714	7250	35	7980	8080	–25.6	Mollusk: <i>Vallonia</i>

<sup>a</sup> AA = NSF-Arizona AMS Laboratory, University of Arizona; Beta = Beta Analytic Laboratory; CAMS = Lawrence Livermore National Laboratory.





**Fig. 3.** Stonehouse Meadow total percent organic matter (%TOM), total carbonate (%TC), and mollusk intervals. %TOM and %TC are smoothed using a three point average. Mollusks represent presence.

this, peats resume with occasional silty or clay rich lenses and the sedimentation rate varies between 0.04 and 0.07 cm yr<sup>-1</sup>. At 495 cm depth sediments abruptly change to inorganic clay, then sand and gravel and sedimentation rates increase to 0.21 cm yr<sup>-1</sup> (~5 yr cm<sup>-1</sup>).

Four zones were identified from the cluster analysis based on changes in both pollen percentages and sediments (%TOM and %TC): Zone 1 (8000–7300 cal yr BP; 690–496 cm), Zone 2 (7300–3000 cal yr BP; 496–250 cm), Zone 3 divided into subzones 3a (3000–2200 cal yr BP; 250–180 cm) and 3b (2200–1750 cal yr BP; 180–160 cm) and Zone 4 (1750 cal yr BP to present; 160–0 cm) (Fig. 3). High resolution %TOM and %TC (continuous samples containing between 5 and 30 years per sample) show a striking pattern of change over time (Fig. 3). In Zone 1, %TOM is low (10–15%) then increases in Zone 2 (average >52%) before declining again in Zone 3 to <10%. Percent total carbonate increases from ~10% in Zone 2 to >30% in Zone 3. In Zone 4 sediments again become organic rich, reaching >80% total organic matter near the surface with low %TC (10–15%).

#### 4.3. Mollusks

Modern mollusks recovered from surface samples in Stonehouse Meadow at the time of coring include the pea clam *Pisidium* sp., and gastropods *Gyraulus* sp., *Physa* sp., and *Euconulus fulvus*. Additional modern data are limited; however Spheriidae (clams), Hydrobiidae (spring snails), and *Physa* (tadpole snails) are considered dominant taxa and *P. kolobensis* occurs in some locations (Southern Nevada Water Authority, 2010). Five molluscan taxa were recovered from core sediments: the aquatic Hydrobiid spring snail, *P. kolobensis*, and four terrestrial taxa *Vallonia albula*, *Euconulus fulvus*, *Discus cronkhitei* and *Vertigo modesta* (Table 2).

*P. kolobensis* is one of almost 100 species of obligatory, spring-dwelling gastropods that occupy high quality, isolated Great Basin wetlands (Hershler and Sada, 1987; Hershler, 1998, 1999). *Pyrgulopsis* are most abundant near spring sources (density often exceeding 10,000/m<sup>2</sup>) where physicochemical environments are stable compared to downstream reaches of spring brooks where seasonal and daily environmental variability is relatively high

**Table 2**

Molluscan taxa recovered from Stonehouse Meadow.

Begin	End	Fragments	Juvenile	Aquatic				Terrestrial			
				<i>Pisidium</i>	<i>Pyrgulopsis kolobensis</i>	<i>Gyraulus</i>	<i>Physa</i>	<i>Vallonia albula</i>	<i>Euconulus fulvus</i>	<i>Vertigo odesta</i>	<i>Discus cronkhitei</i>
Modern		X		X	X	X	X		X		
170	171				X						
175	176	X						X			
180	181	X	X		X			X	X		
185	186	X	X					X			
190	191	X						X			
195	196	X			X			X			
200	201	X			X			X			
205	206	X						X			
210	211	X			X			X			
215	216	X			X						
220	221	X	X		X			X			
225	226	X			X			X			
230	231										
235	236	X			X			X	X		
240	241	X						X		X	X
245	246	X	X		X			X			
250	251	X									
580	581	X			X					X	
585	586										
590	591	X									
595	596	X									
600	601	X									
625	626	X						X			
626	627										
630	631	X									
635	636										
640	641										
645	646							X			
650	651										
655	656										
660	661	X						X			
665	666										

(Hershler, 1998). The presence of *P. kolobensis* requires consistent high quality water and predictability and reliability of spring flow. They do not occur in habitats that are dry or are scoured by floods (Taylor, 1985). Genetic studies indicate the isolation of many hydrobiid species in springs ranges from thousands to millions of years (Hershler et al., 2007; Hershler and Liu, 2008).

The terrestrial mollusks recovered all require moist environmental conditions. Common habitats for these taxa include under leaves and stone (Chamberlin and Jones, 1929), grassland (Bequaert and Miller, 1973), near springs (Henderson, 1936), grass stems and dead leaves at the borders of ponds and marshes (Pilsbry, 1948).

Fig. 3 shows mollusks occur in two intervals, 660–580 cm depth (8000–7600 cal yr BP) and 251–170 cm depth (2800–1850 cal yr BP). Recovery of *P. kolobensis* at 581–580 cm depth indicates that this species was present in Stonehouse Meadow ~7600 years ago. The co-occurrence of the aquatic *P. kolobensis* and the terrestrial *V. albula* in the youngest mollusk interval suggests *V. albula* inhabited a moist, above-water vegetation patch near the orifice. The three terrestrial species *E. fulvus*, *V. modesta*, and *D. cronkhitei*, which occur occasionally in the core, are also likely inhabitants of these near-orifice above-water vegetation patches.

Mollusk occurrence in the Stonehouse cores represents drier intervals. During prolonged drought periods, the spring was likely restricted to a small pool near an orifice. Moist habitat required by terrestrial mollusks would also be restricted during drought, therefore terrestrial mollusks would concentrate in moist areas near a spring orifice and along the banks of spring brooks. The mollusk intervals are concurrent with low %TOM and high %TC and compare well with the cluster analysis zonation of Zones 1 and 3. These drier intervals represent evaporative, alkaline environments

favorable to the precipitation and preservation of carbonate minerals.

Mollusks were likely present during other time periods, but a number of indicators lead us to the conclusion that their aragonite shells were not preserved because of acidification from vegetation decay. First, the coring location had over 30 cm of vegetation mat above sediment indicating that buildup of roots and plant mass is common. Second, the two periods with mollusks present have low percentages of organic matter and periods without mollusks present have high percentages of organic matter. This suggests that root mats may have contributed to increased organic matter (and acidification from their decay) during moist periods. Third, dry conditions are supported by our record of terrestrial molluscan taxa. If the area at the coring location was wetland during these times, few terrestrial taxa would be recovered.

#### 4.4. Diatoms and testate amoebae

Twenty-seven samples were analyzed for diatoms and chrysophytes of which 26 contained siliceous microfossils (Fig. 4). The lowest sample (665–666 cm) was barren. The diatom assemblage throughout the length of the core is dominated by fresh to slightly brackish, alkaliphilic species characteristic of the western desert regions (Potapova and Charles, 2002). The abundance of planktonic taxa (*Aulacoseira* spp., *Cyclotella* spp., *Fragilaria* spp.) is consistently low. Small Fragilarioids (*Pseudostaurosira* spp., *Staurosira* spp., *Staurosirella* spp.) are not present in substantial numbers, except at the top of the core. A large component of the assemblage is comprised of taxa (*Epithemia* spp., *Rhopalodia* spp., *Nitzschia* spp.) that contain nitrogen-fixing cyanobacterial endosymbionts in

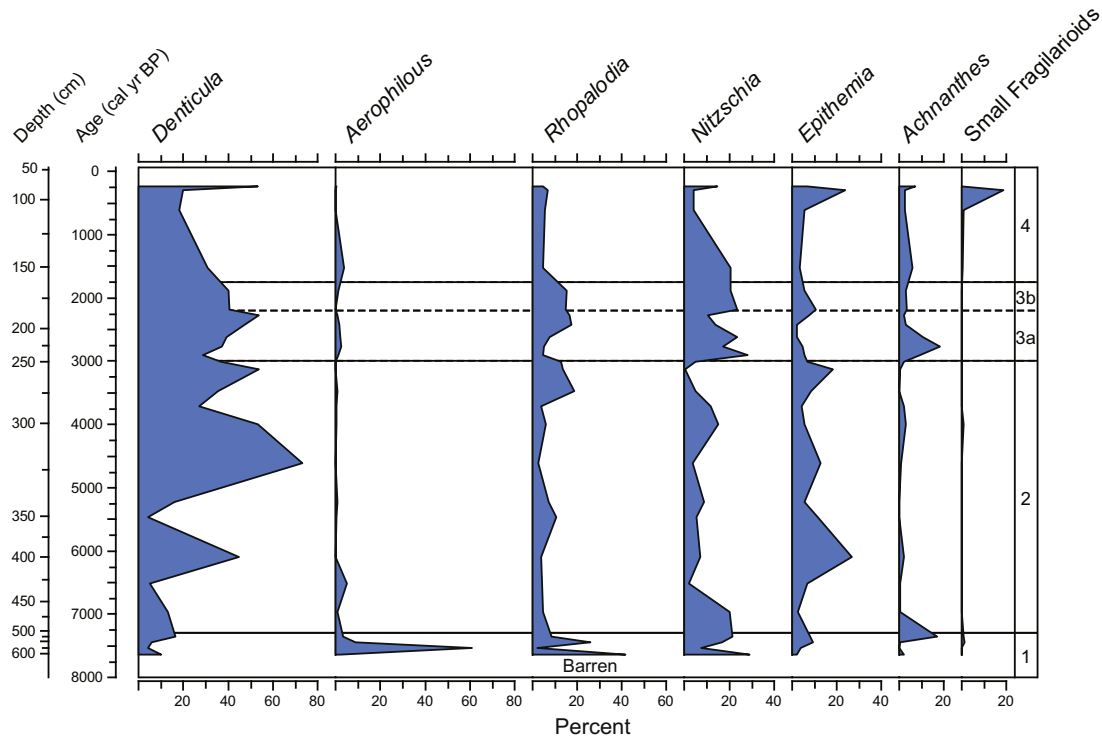


Fig. 4. Relative abundance of selected Stonehouse Meadow diatom genera.

modern environments. Low nitrate levels lead to an increase in the endosymbiont levels. The endosymbiont levels are affected by phosphorous levels, which suggest that the abundance of *Epithemia* spp. and *Rhopalodia* spp. may indicate variations in the amount of available phosphorous in spring water. These taxa are present at Blue Lake warm springs located on the western margin of the Bonneville basin (Kaczmarzka and Rushforth, 1984). The sediments from interval 281–282 cm contain several species (*Euglypha* spp.) of testate amoebae.

The earliest part of Zone 1 (8000–7300 cal yr BP) is dominated by species of *Rhopalodia* (*R. accuminata*, *R. gibberula*, *R. musculus*) and *Nitzschia* (*N. amphibia*, *N. frustulum*) which can survive in environments with elevated levels of total dissolved solids (TDS) often in Na–HCO<sub>3</sub>-dominated systems. This interval is followed by an assemblage dominated by aerophilous taxa (*Hantzschia amphioxys*, *Luticola cohnii*, *L. mutica*, *L. nivalis*, *Pinnularia borealis*) typically found in the western US springs, wet walls, and soils (Johansen et al., 1981, 1984; Flechtner et al., 1998, 2008). This group of species does not appear in substantial numbers in the remainder of the core, suggesting that water flowed from the spring system throughout the Holocene, although the amount of water is variable. Changes in the diatom assemblage (increased *Caloneis* spp., *Achnanthes* spp., *Pinnularia* spp., *Amphora* spp.) in the upper part of Zone 1 suggest an increase in moisture. The abundance of chrysophytes stomatocysts is low during this interval.

Facultative heterotrophic taxa (*Epithemia* spp.) reach maximum abundance in Zone 2 (7300–3000 cal yr BP). *Denticula* spp. and *Nitzschia* spp. also comprise a significant part of the assemblage in this zone, with *Denticula* reaching its highest levels at a depth of 320 cm. This genus is dominant through the remainder of the core, which suggests an increase in TDS. A single peak (48%) in *Caloneis bacillum* occurs at a depth of 350 cm. Periphytic species, including those capable of withstanding higher levels of TDS increase in abundance in the upper part of this interval. Aerophilous species

account for less than 5% of the assemblage and continues at this level of abundance up core.

*Denticula* spp. comprise between 40 and 50% of the assemblage in Zone 3 (3000–1750 cal yr BP). *Epithemia* spp. and *Rhopalodia* spp. decrease in abundance in Zone 3a (3000–2200 cal yr BP) but return to previous abundance levels in Zone 3b (2200–1750 cal yr BP). These taxa are replaced in Zone 3a by periphytic species (*Nitzschia* spp., *Navicula* spp., *Gomphonema* spp.) capable of surviving at higher levels of TDS. The increased abundance of *Nitzschia* spp. (*N. amphibia*, *N. frustulum*) also indicates an increase in TDS. *Achnanthes* spp. reach an isolated peak of 18% in the middle of Zone 3a, indicating a possible short-term interval of higher moisture. There is a slight increase in the abundance of epiphytic species in Zone 3b.

Zone 4 (1750 cal yr BP to present) is characterized by an increase in *Denticula* spp. and freshwater species (*Achnanthes* spp., *Pinnularia* spp.). Small fragiliarioids comprise almost 20% of the assemblage at the top of the core, indicating higher water levels.

#### 4.5. Pollen

Zone 1 (8000–7300 cal yr BP) is dominated by high percentages of shrubs (Amaranthaceae and Rosaceae) and trees with maximum percent *Pinus* ~7300 cal yr BP (Fig. 5). Pollen accumulation rates are highest in this zone. Pollen accumulation rates are low through most of the core, but the pattern of variation is consistent with the pollen percentage data.

Zone 2 (7300–3000 cal yr BP) is characterized by high percentages of trees (*Pinus* and *Juniperus*) and shrubs (*Artemisia*, Amaranthaceae and *Sarcobatus*). Cyperaceae accumulation rate is high, indicating the establishment and persistence of the wet meadow.

In Zone 3 (3000–1750 cal yr BP), grass (Poaceae) increases from ~10% to 30%, members of the sunflower family (Asteraceae) increase from 5% to 30% and sedge (Cyperaceae) accumulation rate

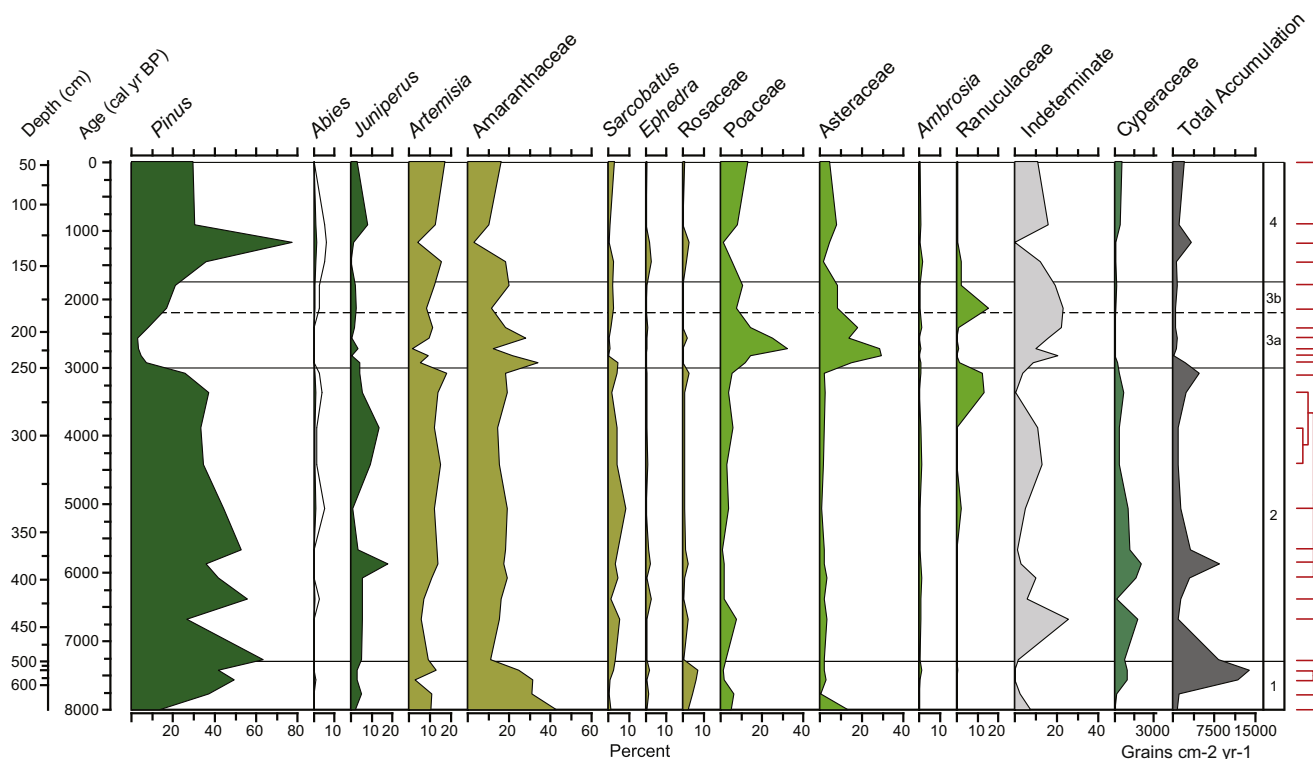


Fig. 5. Stonehouse Meadow pollen percentages, Cyperaceae pollen accumulation rates and total pollen accumulation rate. Unfilled black line represents 5× exaggeration.

decreases. Desert saltbush plants (Amaranthaceae) increase while trees (*Pinus* and *Juniperus*) decline markedly. Changes are most pronounced in Zone 3a (3000–2200 cal yr BP) with Zone 3b (2200–1750 cal yr BP) being somewhat transitional with a decrease in drought-tolerant grasses and herbs, an increase in trees, increasing %TOM and decreasing %TC. An interesting feature of the transition before and after Zone 3a is the presence of Ranunculaceae, possibly *Thalictrum* (meadow rue), apparently invading the meadow.

Zone 4 (1750 cal yr BP to present) marks a return of trees. Poaceae and Asteraceae persist, although at lower percentages, and Cyperaceae returns, although not in the abundance seen prior to 3000 cal yr BP.

## 5. Discussion

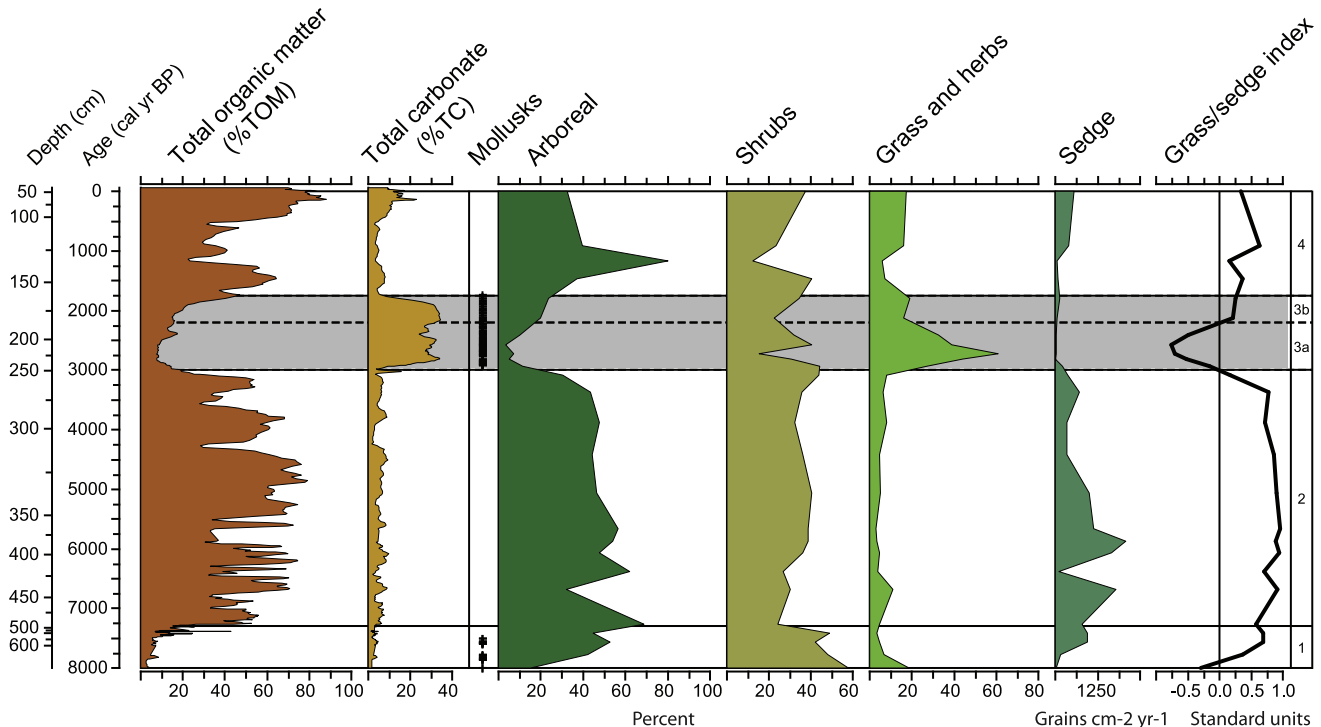
### 5.1. Early and middle Holocene climate in the eastern Great Basin

In the eastern Great Basin, Holocene climate shifted from cool and moist to warmer and drier sometime between about 9500 and 9000 cal yr BP (Grayson, 2000; Schmitt et al., 2002). This shift is interpreted from the decline or extinction of small mammals adapted to cool and moist habitats, such as yellow bellied marmot, western harvest mouse, and pygmy rabbit, and an increase in dry adapted species such as kangaroo rat and ground squirrels, seen in changes in the abundance of bones recovered from Homestead Cave (Madsen et al., 2001), Camel Back Cave (Schmitt et al., 2002) and the Bonneville Rockshelter (Schmitt and Lupo, 2012) in the Bonneville Basin. Pollen and sedimentary evidence from Blue Lake on the western margin of the Bonneville Basin show a decline in marsh taxa and increase in playa-margin shrublands following 9800 cal yr BP, with the potential desiccation of Blue Lakes ~8000 cal yr BP (Louderback and Rhode, 2009).

Stonehouse Meadow was either absent or severely restricted in size at 8000 cal yr BP. This is supported by high percentages of saltbush, low %TOM, and high numbers of aerophilic diatom taxa typical of subaerial environments. Beginning ~7500 cal yr BP, the meadow expanded, seen by the appearance of peaty sediments, moisture dependent diatoms, and sedges. The decline in saltbush indicates loss of shrubland, probably from the valley floor, as the meadow grew. Once a wet meadow became established, by 7300 cal yr BP, it persisted. Emergence of the meadow suggests a shift to wetter conditions following the period of early Holocene aridity. The Great Salt Lake may have been almost completely dry ~7800 cal yr BP (Madsen et al., 2001) but had risen again by ~6700 cal yr BP suggesting increased effective moisture sometime after 7000 cal yr BP. This is consistent with vegetation changes at the eastern edge of the Great Basin that indicate wetter climate by ~6800 cal yr BP (Madsen and Currey, 1979). Louderback and Rhode (2009) record a decrease in Amaranthaceae pollen in relation to *Artemisia* between 7000 and 6500 cal yr BP which they interpret as a shift towards cooler temperatures and wetter climate. In contrast, at Ruby Marsh, pollen evidence for an increase in Amaranthaceae pollen ~7800 cal yr BP is interpreted as a shift to warmer and/or drier conditions (Thompson, 1992).

Pinyon pine migrated north during the early middle Holocene from the Mojave Desert Pleistocene refugium into the eastern Great Basin (Grayson, 2011). The increase in pine pollen at Stonehouse Meadow by 7800 cal yr BP and peak abundance at 7300 cal yr BP is consistent with an increase in pine pollen ~7500 cal yr BP at Blue Lakes (Louderback and Rhode, 2009) and the first appearance of pinyon pine in packrat middens at Danger Cave by 7580 cal yr BP (Rhode and Madsen, 1998). It has been argued that humans may have introduced pinyon pine nuts to the Danger Cave site (Cole et al., 2012), but our pollen record indicates that pine was well established in the region by this time.





**Fig. 6.** Stonehouse Meadow summary diagram. Arboreal pollen includes *Pinus*, *Juniperus* and *Abies*. Shrub pollen includes *Amaranthaceae*, *Artemisia*, *Sarcobatus*, *Ephedra*, and *Rosaceae*. Grass and herbs include *Poaceae*, *Asteraceae* and *Ambrosia*. Sedge (*Cyperaceae*) is plotted as pollen accumulation rate. The grass/sedge index is calculated using the formula  $[(\text{Poaceae} + \text{Cyperaceae})/(\text{Poaceae} - \text{Cyperaceae})]$ . Gray shading represents the proposed Late Holocene Dry Period.

There are no large changes in vegetation evident in the Stonehouse Meadow pollen record in Zone 2 between ~7300 and 3000 cal yr BP (Fig. 4). Pinyon/juniper woodland and sagebrush steppe dominated the surrounding landscape, and sedge generally persisted in the meadow. The presence of testate amoebae (*Euglypta*) indicates the presence of wet conditions and possibly standing water (Woodland et al., 2008). Several regional records suggest a shift towards greater effective moisture in the eastern Great Basin beginning sometime between 4800 and 4400 cal yr BP, but not all of these records are well dated through this interval, and the sampling resolution is low for some studies. Louderback and Rhode (2009) suggest a shift to predominantly cooler and wetter conditions between ~4400 and 3400 cal yr BP based on increases in pine and sagebrush at Blue Lakes. A shift towards greater effective moisture is also suggested at Ruby Marsh after ~4800 cal yr BP interpreted from a steady increase in pine pollen and decrease in *Amaranthaceae* (Thompson, 1992). Mehringer (1985) identified a shift to wetter conditions at Crescent Spring near the Great Salt Lake beginning about 3800 cal yr BP, although the error associated with this age date is very large and Mehringer notes that interpretation of the record is complicated by rebound and flooding of the Great Salt Desert.

## 5.2. Evidence for a Late Holocene Dry Period (~2800–1850 cal yr BP)

We infer a period of prolonged drought at Stonehouse Meadow in Zone 3, between ~2800 and 1850 cal yr BP, interpreted from the decline of sedge meadow, loss of organic material, decline in trees and shrubs, increase in grasses and herbs, increase in total inorganic carbon, changes in diatom species, and preservation of mollusks (Fig. 6). Castelli et al. (2000) described four meadow types in the riparian zones of central Nevada based on two criteria, the

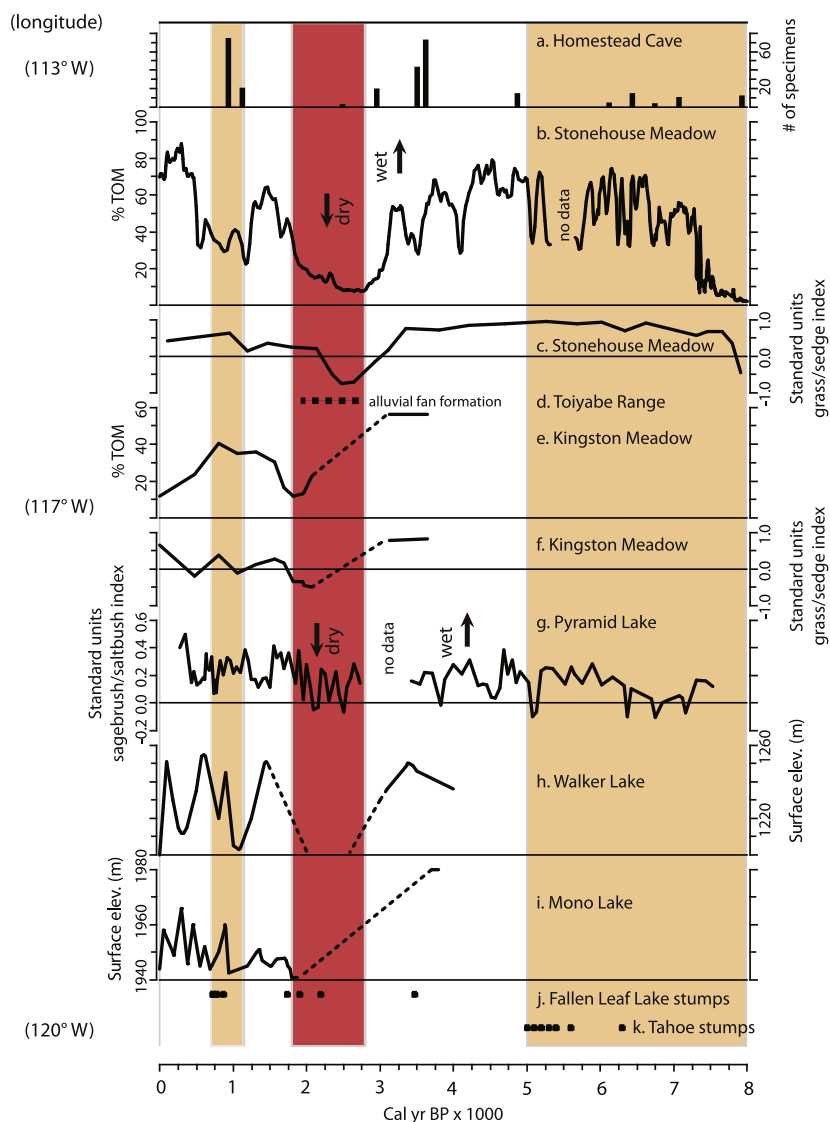
presence of sedges or grass and depth to water table. Both sedge and grass can occur in all four meadow types, but sedges require standing water or a water table near the surface. As the water table declines, sedges decline and grasses increase (Castelli et al., 2000). We interpret the decline in sedge and increase in grass as a prolonged period (ca 1000 years) of lowered water table.

We argue that there is evidence from multiple different proxies across the Great Basin for an extended period of persistent drought between ~2800 and 1850 cal yr BP. We refer to this period as the Late Holocene Dry Period. Here we review the current evidence for this dry period and begin to define its geographic extent, duration and magnitude across the Great Basin (Fig. 7).

### 5.2.1. Eastern Great Basin and Bonneville Basin

The Bonneville Basin lies to the east of Stonehouse Meadow (Fig. 1) in the eastern Great Basin within the state of Utah. A number of studies of the Great Salt Lake and its environment have examined changes in lake level in the late Holocene. A careful review of these studies is required to determine the timing of the initiation of drought within the Bonneville Basin.

Madsen et al. (2001) stated that between ~3100 and 2450 cal yr BP the Bonneville Basin experienced one of the most dramatic climatic cycles of the Holocene, with a major cooling event, a decline in treeline in some areas, and expansion of the Great Salt Lake to an elevation of ~1287 m, flooding the Great Salt Desert as far west as the Nevada border. They acknowledge that chronologies for many of the studies during this period are poor, and cite the Crescent Spring pollen study (Mehringer, 1977, 1985) for age control. Crescent Spring, Utah, is a playa-edge spring-fed salt marsh along the western edge of the Great Salt Lake (Mundorf, 1971). Mehringer (1977) noted that stratigraphy, fossil pollen and diatoms led to an interpretation of probable flooding of the Great Salt desert between ~3800 and 2200 cal yr BP. Pollen evidence from Crescent



**Fig. 7.** Stacked data of sites across the central Great Basin from east to west that record an extended dry period between 2800 and 1850 cal yr BP. Light orange bands represent established periods of extended dry climate, the Medieval Climate Anomaly (~1150–700 cal yr BP) described by Stine (1994) and others, and the Middle Holocene from about 7500 to 5000 cal yr BP (Grayson, 2011). Red band represents the Late Holocene Dry Period (~2800–1850 cal yr BP) described in this paper. Data include a) Number of fish bones (*Gilia atraria*) deposited by owls in cave deposits (Broughton et al., 2000; Madsen et al., 2001; Terry, 2008); b) Percent total organic matter (%TOM) in Stonehouse Meadow (this study); c) Grass/sedge index for Stonehouse Meadow calculated as  $[(\text{Poaceae} + \text{Cyperaceae})/(\text{Poaceae} - \text{Cyperaceae})]$  (this study); d) Timing of alluvial fan construction in the Toiyabe, Toquima and Monitor Ranges (Miller et al., 2004); e) Percent total organic matter (%TOM) in Kingston Meadow (Smith, 2003; Mensing et al., 2007); f) Grass/sedge index for Kingston Meadow calculated as  $[(\text{Poaceae} + \text{Cyperaceae})/(\text{Poaceae} - \text{Cyperaceae})]$  (Smith, 2003; Mensing et al., 2007); g) Sagebrush/saltbush index for Pyramid Lake calculated as  $[(\text{Artemisia} + \text{Amaranthaceae})/(\text{Artemisia} - \text{Amaranthaceae})]$  (Mensing et al., 2004); h) Surface elevation of Walker Lake (Adams, 2007); i) Surface elevation of Mono Lake (Stine, 1990; Davis, 1999); j) age dates for submerged stumps in Fallen Leaf Lake (Kleppe et al., 2011); k) age dates for submerged stumps in Lake Tahoe (Lindström, 1990). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Spring shows a shift beginning ~3800 cal yr BP towards increased abundance of *Artemisia* (sagebrush) in relation to *Amaranthaceae* (halophytic salt desert shrubs) (Mehring, 1985). The Crescent Spring site has three radiocarbon ages on bulk sediment using standard radiocarbon methods ( $^{14}\text{C}$  yr 4350  $\pm$  120; 3490  $\pm$  300; and 2200  $\pm$  50). To our knowledge, these are the only radiocarbon ages that support an interpretation of higher lake levels for the Great Salt Lake between 3800 and 2200 cal yr BP.

Currey (1990) found beach deposit evidence for a Great Salt Lake highstand of 1286.7 m above sea level (~6.7 m above modern), but the radiocarbon ages of shoreline features only constrained this event between 7900 and 1300 cal yr BP. He suggested that the highstand (also see Fig. 3 in Currey and James, 1982) occurred

between ~3200 and 2000 cal yr BP based on earlier studies, including Mehring (1985) noted above, and the oxygen isotope study of McKenzie and Eberli (1987). McKenzie and Eberli (1987) presented oxygen isotope evidence to support the presence of a late Holocene highstand, but their only age control was the Mazama ash, dated to ~7600 cal yr BP (Hallett et al., 1997). They found Mazama ash and correlated mineralogic changes above the ash with the pollen record presented in Mehring (1985).

Grayson (2011) echoes the statement that the highest stand of the Great Salt Lake in the Holocene occurred in the period between ~3600 and 2000 cal yr BP, but goes on to note that Madsen et al. (2001) has evidence that the lake had fallen by ~2450 cal yr BP. Small animal remains from stratified owl deposits in Homestead

Cave, Utah, were analyzed as a proxy for climate change (Madsen et al., 2001). Owls have a foraging range of ~5 km and the number of fish bones from the native Utah chub (*Gilia atraria*) was used as a proxy for expansion and contraction of the lake, with high numbers of bones interpreted as evidence that lake level was higher (within 5 km of the site) and absence of bones interpreted as low lake levels (farther than 5 km from the site). The number of fish bones peaked in stratum XII between ~3700 and 3570 cal yr BP (Broughton et al., 2000) then declined steadily, reaching a minimum in stratum XV which had almost no fish bones (Fig. 7). Recent age control on kangaroo rat (*Dipodomys* sp.) bones in stratum XV found that 17 of 20 radiocarbon ages fell between 3250 and 1825 cal yr BP (Terry, 2008). Three ages were sufficiently different to suggest some mixing of the sediments. The modal age for these dates was 2575 cal yr BP. A date on artiodactyl pellets from stratum XIV (below XV) had an age of 2970 cal yr BP (Broughton et al., 2000).

All previous studies that cite high lake levels for the Great Salt Lake until 2200 cal yr BP rely on one standard radiocarbon age taken on bulk sediment from the Crescent Spring pollen study first reported in Mehringer (1977). There have been no other independent confirmations of this age for a Holocene Great Salt Lake highstand. The multiple AMS radiocarbon ages recently obtained on small mammal bones from stratum XV in Homestead Cave as well as the age dates on the deeper stratum XIV suggest that sometime after 2900 and probably by 2600 cal yr BP, the Great Salt Lake had retreated from its Holocene highstand elevation of 1286.7 m. We argue that the well-dated Homestead cave site supports an interpretation of low lake levels at the Great Salt Lake beginning ~2600 cal yr BP and possibly continuing until 1825 cal yr BP. Low lake levels between 2900 and 1600 cal yr BP were inferred from diatom evidence at Bear Lake, in the eastern Great Basin within the Great Salt Lake watershed (Moser and Kimball, 2009). Similarity between the historic Bear Lake and Great Salt Lake hydrographs suggest that the hydrology of Bear Lake reflects regional precipitation.

Lake Sevier in the Sevier Basin, a sub-basin of the Bonneville Basin, experienced a highstand to 1382.3 m about 2620 cal yr BP, based on a  $^{13}\text{C}$  adjusted radiocarbon date of  $2560 \pm 75$  on gastropod shells found in beach deposits (Oviatt, 1988). No age is given for the subsequent lake regression, however given no younger age for the highstand, one can assume that the lake retreated following this date, which is consistent with our interpretation of the Great Salt Lake data.

Treeline evidence from Mount Washington in the Snake Range recorded an expansion of bristlecone treeline at least 100 m higher than modern between ~4400 and 2200 cal yr BP (LaMarche and Mooney, 1972). Today trees at the upper elevational extent are stunted (krumholz) forms, but remnant wood suggests these late Holocene trees were upright. Physiological studies suggest that one of the triggers for cold-hardiness that allows upright forms to persist through the winter is summer drought the previous year (Alden and Hermann, 1971). Evidence of upright trees above the modern treeline limit is inferred as a period of drier and warmer climate between ~4400 and 2200 cal yr BP (LaMarche and Mooney, 1972).

### 5.2.2. Central Great Basin

Geomorphic evidence from the Toiyabe, Toquima and Monitor Ranges also supports the interpretation of a dry period between 2600 and 1900 cal yr BP (Fig. 7). An episode of major alluvial fan building occurred synchronously across multiple watersheds between ~2600 and 1850 cal yr BP, suggesting that aggradation was initiated by a regional change in the environment (Miller et al., 2004). They infer that the period of intensive aggradation was

caused by a shift from wet to dry climate associated with a significant change in the upland vegetation. Decreased density of vegetation during extended dry climate resulted in increased runoff and hillslope erosion during precipitation events initiating an episode of side-valley alluvial fan building and valley bottom aggradation.

Kingston Meadow in the Toiyabe Range of central Nevada shows a similar pattern to Stonehouse Meadow (Fig. 7) being abruptly transformed from a wet sedge meadow to a dry grassy meadow by at least 2030 cal yr BP (209 cm depth) and persisting for a period afterward (Mensing et al., 2007). Sediment at 242 cm depth age dates to 3800 cal yr BP. An unconformity at 210 cm depth suggests the possible loss of sediments between 2030 and 3800 cal yr BP, preventing the determination of onset of drought from this record. Data from woodrat middens adjacent to Kingston Meadow and nearby riparian areas show a dramatic drop in number of plant taxa ~2500 cal yr BP, suggesting reduced stream flow and a potential beginning age for the drought at this site (Tausch et al., 2004).

### 5.2.3. Western Great Basin

Extensive evidence for a persistent dry period between ~2800 and 1850 cal yr BP comes from pollen, lake levels, and treeline data in the western Great Basin. Pollen evidence from Pyramid Lake showed that between 2500 and 2000 cal yr BP the ratio of sagebrush to saltbush pollen (Fig. 7) was as low as during the middle Holocene suggesting low lake levels associated with drought (Mensing et al., 2004). Saltbush grows in drier habitats than sagebrush and also tolerates saline playas, suggesting expansion of saltbush onto exposed lake sediments during periods of low lake levels. Minima in the pollen ratios correspond with oscillations in the  $\delta^{18}\text{O}$  record interpreted as droughts (Benson et al., 2002).

Multiple lines of evidence suggest a lowstand at Walker Lake (Fig. 7) between ~2800 and 2100 cal yr BP (Bradbury et al., 1989; Benson et al., 1991). The presence of *Ruppia* pollen in sediment cores suggests saline water with a depth of <1 m (Bradbury, 1987; Bradbury et al., 1989). The absence of freshwater diatoms (*Stephanodiscus*) and the presence of euryhaline *Navicula* spp. suggest that Walker Lake was a shallow saline lake at this time. High percentages of greasewood (*Sarcobatus*) pollen indicate expansion of drought tolerant species in the landscape and probably onto newly exposed playa surfaces. Very low lake levels between ~2700 and 2400 cal yr BP are interpreted from  $\delta^{18}\text{O}$  data in a sediment core from Walker Lake (Yuan et al., 2006). Dates on tufas at 1211–1212 m elevation date to between 2200 and 1740 cal yr BP, confirming that the lake was still very low during this period (Adams, 2007). Recent age control on organic materials collected from fluvial, deltaic, and beach sediments in geomorphic settings around the lake placed the beginning of the regression sometime after 3220 to 2950 cal yr BP and the end of the lowstand sometime prior to 1550 to 1420 cal yr BP (Adams, 2007).

At Mono Lake, the Marina Recession and Low Stand is provisionally dated at 1810 cal yr BP (Stine, 1990). The increased abundance of the diatom *Anomoeneis sphaerophora* indicates an increase in salinity during this time (Zimmermann et al., 2012). The lake fell to 1940.9 m elevation, as low as any level in the late Holocene (Fig. 7). Although the timing of the initiation of the recession is uncertain, sand lenses and increased *Artemisia* pollen from lake sediment cores date to 2400 cal yr BP (Davis, 1999) indicating that the lake was already low by then.

Submerged rooted tree stumps in Fallen Leaf Lake at the southern edge of Lake Tahoe indicate a low lake between ~2320 and 1620 cal yr BP (Kleppe et al., 2011). Upright stumps a full 35 m below the present surface of the lake have been radiocarbon dated to 2320–2060, 1990–1820 and 1860–1620 cal yr BP (Fig. 6). Kleppe et al. (2011) argue that precipitation must have dropped to or below

60% of normal for at least 100 years to lower the lake 35 m. The record of low lake levels across the western Great Basin coincident with the timing of these submerged tree stumps suggests that this was a period of sustained drought.

At Lead Lake in the Carson Sink higher values of *Typha latifolia* (broadleaf cattail) and *Typha angustifolia* (narrowleaf cattail) between 2200 and 1850 cal yr BP indicate the presence of a shallow marsh and influx of water (Wigand and Rhode, 2002). But, increases in *Potamogeton* (pondweed) ~1850 cal yr BP indicate deeper water beginning at that time. Grayson (2011) suggested that the Lead Lake record was potentially evidence for a wet climate in the western Great Basin at the same time that deep closed basin lakes were low. Lead Lake is a shallow system, and small changes in temperature or inflow could have resulted in large ecologic changes. The presence of a shallow marsh, but not water sufficiently deep to support *Potamogeton* suggests only a modest influx of water at this site between 2200 and 1850 cal yr BP.

#### 5.2.4. Paleotemperature

Paleotemperature reconstructions for the Great Basin during the late Holocene are limited. Late Holocene glacial advances have been reconstructed for the Palisade Glacier in the eastern Sierra Nevada by dating sediment cores with glacial outwash (Bowerman and Clark, 2011). They identified a reinitiation of glacial advances beginning ~3200 cal yr BP with a possible glacial maximum ~2800 cal yr BP and a distinct glacial maximum at ~2200 cal yr BP. Cooling was estimated at 0.2–2 °C with potential increased moisture of 3–26 cm snow water equivalent. A high elevation record of foxtail pine at nearby Cirque Peak found that many trees died between 2500 and 2300 yr BP (absolutely dated dendrochronological records) and timberline fell ~25 m (Scuderi, 1987). Narrow ring growth patterns suggest that the tree death and lowered timberline were associated with colder temperatures.

Across the Great Basin at Stella Lake in the Snake Range, Reinemann et al. (2009) reconstructed a paleotemperature record using chironomids from lake sediments. The chironomid-inferred mean July temperature reconstruction suggested a late Holocene temperature minimum between 3000 and 2000 cal yr BP of ~1.5 °C below modern. In contrast, remnant bristlecone pine treelines on Mount Washington, also in the Snake Range, grew 100 m above modern elevational limits between ~4400 and 2200 cal yr BP (LaMarche and Mooney, 1972). Those authors interpreted this period as warmer summer temperatures than modern. Additional paleotemperature studies are needed to further develop our understanding of temperatures during this period.

#### 5.3. Geographic extent of the Late Holocene Dry Period

We have presented multiple lines of evidence from a diverse set of proxy records extending from the eastern Sierra Nevada across the central Great Basin to the Great Salt Lake that support extended drought conditions that we term the Late Holocene Dry Period. Some evidence suggests that this dry period began as early as 2800 cal yr BP, and nearly all records support a dry climate between 2600 and 1850 cal yr BP. In contrast, sites in the northern Great Basin record either no clear evidence of drought, or have wetter than average climate during this period, suggesting that the northern boundary between wet and dry climates may have been between about 40° and 42° N latitude.

##### 5.3.1. Northern Great Basin

Diamond Pond and the Malheur Marshes of the Harney Basin in southern Oregon had deeper water reflecting more effective moisture between 2900 and 2000 cal yr BP (Wigand, 1987). Evidence for wetter climate at this time is supported by increases in

aquatic plant seeds as well as increased *Juniperus* pollen, suggesting the spread of woodland into shrub communities. Pollen evidence from Wildhorse Lake in the Steens Mountains supports the Diamond Pond record of increased effective moisture between 3000 and 2000 cal yr BP evidenced by increasing Poaceae in relation to *Artemisia* pollen (Mehring, 1985). Similarly, expansion of pine at Craddock Meadow, Oregon, is interpreted as a period of moister climate (Wigand, 1989). Wigand and Rhode (2002) argued that winter precipitation increased relative to summer precipitation.

Patterson Lake, Lilly Lake and Dead Horse Lake in the northwestern Great Basin show no clear pattern of change from pollen and charcoal data during the period from ~3000 to 2000 cal yr BP (Minckley et al., 2007). The pollen records are dominated by *Pinus* and *Artemisia* and show little variation. Fire histories show an extended period of low fire episode frequency between 3000 and 2200 cal yr BP at Lilly Lake, and decreasing fire frequency between 2800 and 2000 cal yr BP at Patterson Lake. During this same period Dead Horse Lake has an increase in fire episode frequency reaching a maximum at 2000 cal yr BP. Reconstructed annual precipitation anomalies also show no clear trends. A low-resolution pollen study from Bicycle Pond has very few data points during this period, but records an increase in *Pseudotsuga* (douglas fir) pollen, a species that requires more precipitation than does pine (Wigand and Rhode, 2002).

A high resolution pollen study (samples interval ~55 years) from Mission Cross Bog (2424 m elev., 41.8° N lat.) in the Jarbidge Mountains within the headwaters of the Humboldt River drainage in northeastern Nevada shows variable climate between 2800 and 1800 cal yr BP (Mensing et al., 2007). Although named a bog, it is a fen maintained throughout the year by groundwater. The fen is located near the sagebrush/montane forest ecotone and the ratio of conifer (*Pinus* and *Abies*) to sagebrush (*Artemisia*) pollen was used as a measure of drought. Overall the period is wet, but also records three dry periods of up to a century in duration centered on 2500, 2200 and 1900 cal yr BP. Further to the south, evidence from the Ruby Marsh (40.2° N lat.) provides no clear evidence for either a wetter or drier climate during this period (Thompson, 1992).

##### 5.3.2. Southern Great Basin and Northern Mojave Desert

There is evidence from several different proxies for drier than modern climate in the southern Great Basin between ~2800 and 2300 cal yr BP, however after that date the records become more variable. The Pahrnatagat Lake pollen record used the ratio between conifers and saltbush to interpret periods of wetter and drier climate than modern (Wigand, 1997; Wigand and Rhode, 2002). The periods from 2700 to 2300 and 2200 to 2100 cal yr BP were interpreted as drier than modern with a possible short wet period between 2300 and 2200 cal yr BP. From ~2100 to 1850 cal yr BP wetter climate than modern was present.

North of Las Vegas, Nevada, black mats, organic rich layers in desert alluvial deposits associated with spring discharge, were essentially absent from the southern Great Basin between 7000 and 2350 cal yr BP (Quade et al., 1998). One radiocarbon age of 2350 cal yr BP was obtained from one mat at the Valley Wells site (Quade et al., 1998) although there is no evidence for widespread reactivation of springs across the southern Great Basin at this time. The next subsequent black mat recorded from the Las Vegas Valley dated to 1360 cal yr BP (Haynes, 1967). Due to bioturbation, determining accurate dating for black mats is challenging and it is not possible to determine the beginning or ending ages of black mat formation from an individual date. These dates simply represent potential minima for the reinitiation of black mat formation. The Pahrnatagat Lake study and black mat study suggest that there was a return to wet conditions sometime after 2300 cal yr BP, however the exact timing for this transition is not well constrained and the greater influence of summer monsoonal precipitation in the

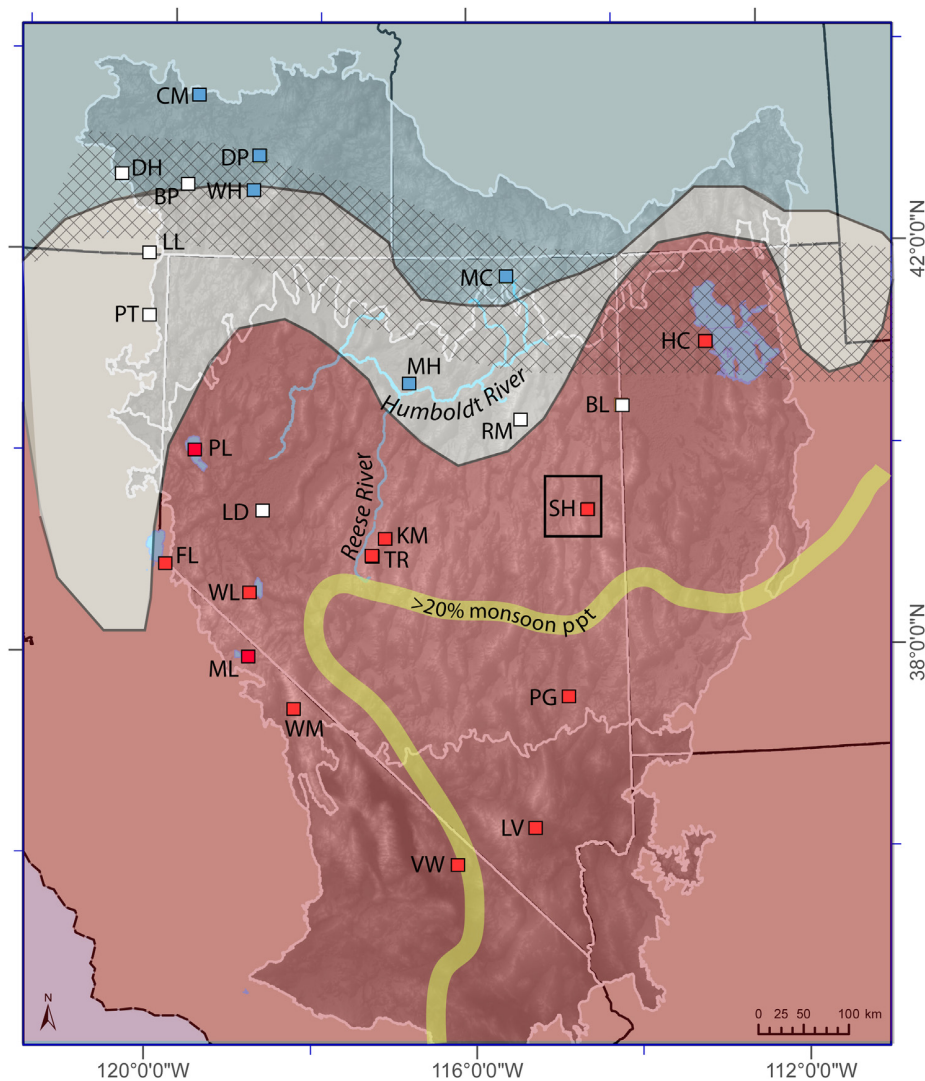


southern Great Basin (Fig. 8) further complicates interpretation of records from this region.

Treeline evidence from the southern Great Basin region shows distinct declines in treeline between 2800 and 2300 cal yr BP, but climatic interpretations for these changes differ. Bristlecone pine treeline on Campito Mountain in the White Mountains moved downslope about 30 m between 2800 and 2500 cal yr BP (LaMarche, 1973). LaMarche (1973) argued that precipitation is a limiting factor on Campito Mountain and treeline changed in response to drier conditions. On nearby Sheep Mountain, treeline did not change, but LaMarche (1973) argued that this site is largely controlled by temperature. In the southern Sierra Nevada at Cirque Peak, 60 km southwest of the White Mountains, foxtail pine died and treeline declined between 2500 and 2300 years ago (Scuderi, 1987), but tree mortality was attributed to colder than modern temperatures, not increased aridity.

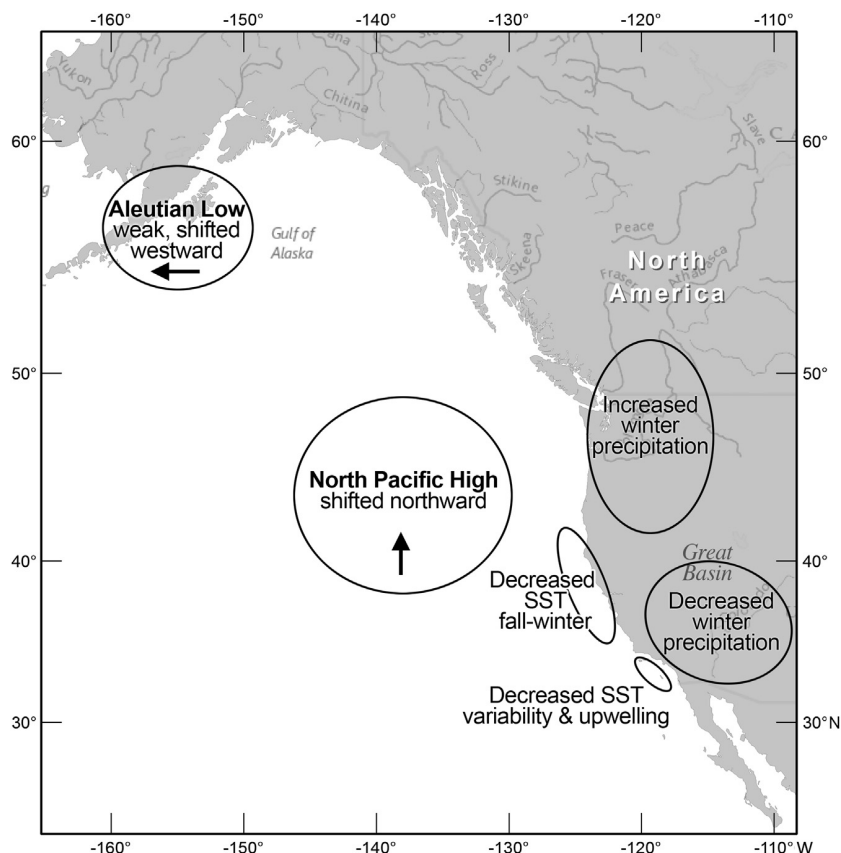
#### 5.4. Spatial climate patterns – western North America

Our evidence suggests a wide-scale and persistent dry period across the central Great Basin occurred between ~2800 and 1850 cal yr BP, supported by a wide range of proxy climate indicators including pollen, macrofossils in packrat middens, faunal remains, geomorphology, and lowered lake levels. Sites in the northern Great Basin, however, either show no clear evidence of drought during this period, or suggest wetter than modern climate. Sites in the southern Great Basin confirm a period of dry climate at least until ~2300 cal yr BP, after which the signal becomes more variable. The evidence suggests a pattern with a dry southwest and wet northwest during this time period. This dry in the southwest and wet in the northwest precipitation record across the Great Basin is supported by large-scale spatial climate pattern reconstructions.



**Fig. 8.** Spatial distribution of sites recording dry climate (red squares), wet climate (blue squares) or no definite signal (white squares) across the Great Basin between 2800 and 1850 cal yr BP. Site names as in Fig. 1. Sites in the southern Great Basin return to a wetter climate ~2300 to 2200 cal yr BP. The wide gray band (hatched lines) represents the boundary defined by Mitchell (1976, Fig. 3) separating the northern Region II, characterized by frequent winter Pacific air masses and interior summer air masses, from southern Region V, characterized by infrequent winter Pacific air masses and interior summer air masses. Narrow yellow band represents the approximate boundary separating those areas in the southern Great Basin that receive >20% of the annual precipitation during the monsoon in July and August (data from WRCC <http://www.wrcc.dri.edu/>). Shading is a generalization of dipole precipitation patterns associated with a positive SOI (adapted from Wise, 2010). Red shading: negative precipitation anomalies; blue shading: positive precipitation anomalies; white shading: transition zone between ~40 and 42° N latitude.





**Fig. 9.** Proposed late Holocene winter spatial climate patterns, adapted from Barron and Anderson (2011) showing general position of air masses, SST and upwelling for dry conditions in the Great Basin and wet conditions in the Pacific Northwest.

First, Mitchell (1976, see Fig. 3) identified air mass boundaries in the western United States based on equivalent potential temperature. Two winter patterns occur: a northwest–southeast gradient across eastern Montana and an east–west gradient along the northern borders of California, Nevada and Utah (Fig. 8). This winter gradient is in the same 40–42° N latitude band as reported by Wise (2010) noted below.

Second, analyses of El Niño–Southern Oscillation (ENSO) related precipitation variability in the western United States identified a north–south pattern of variation between the Pacific Northwest and arid Southwest (Redmond and Koch, 1991; Dettinger et al., 1998). During years with a positive Southern Oscillation Index (SOI), La Niña-like conditions, the storm track moves north, a weak Aleutian low develops with lower pressure over the Western U.S. and the southern region of the dipole experiences anomalously dry climate and the northern region is anomalously wet (Cayan and Redmond, 1994). McCabe et al. (2008) analyzed positive and negative phases of the Pacific Decadal Oscillation (PDO) and Atlantic Multidecadal Oscillation (AMO) and compared them to drought frequency. A negative phase of the PDO is characterized by a shallow Aleutian low, warm western and central North Pacific, and cool eastern Pacific coastal waters and anomalous cooling in the central and eastern tropical Pacific (Gershunov and Barnett, 1998). During a negative PDO, La Niña-like conditions, and positive AMO, high drought frequency can occur in the southwestern U.S. (McCabe et al., 2008). Drought frequency occurs to a lesser extent with negative PDO and negative AMO regimes. More recently a persistent narrow transitional boundary in the dipole has been identified within the 40–42° N latitude band across northern California and

Nevada (Wise, 2010). This pattern is strengthened when either the PDO or AMO are in a negative phase, although it also occurs during the positive phase of the PDO and AMO (Wise, 2010). The dipole pattern is not characteristic of neutral SOI years (Wise, 2010).

Third, Barron and Anderson (2011, Fig. 5) explored late Holocene teleconnections among sites in the eastern Pacific Ocean and its margins. Based on their conclusions, long-term central Great Basin drought could occur with a weakened, westward-shifted Aleutian low, a stronger, northward shifted North Pacific High, decreased summer and fall coastal upwelling off California, and decreased fall and winter sea surface temperatures (SST) along the California (Kennett et al., 2007) and Oregon coasts, compared to modern day conditions (Fig. 9).

Sites across the central Great Basin that record extended dry climate between 2800 and 1850 cal yr BP are south of the dipole transitional zone and spatially within those regions that receive anomalously low winter precipitation during positive SOI years (Fig. 8). Sites in the northern Great Basin fall within the region that is along the transition zone or north, within the area that receives anomalously high winter precipitation.

Miller et al. (2004) described seemingly contradictory data for the Humboldt River that may support the suggestion of a persistent dipole pattern with a dry southwest and wet northwest separated by a transition zone between ~40 and 42° N. They inferred dry climate between 2600 and 1800 cal yr BP during a period of alluvial fan construction in the Toiyabe Range (described above). The western slope of the Toiyabe Range drains into the Reese River, the southernmost tributary of the Humboldt River watershed (Fig. 1). During nearly the same time period, between 3190 and 2080 cal yr

BP, they inferred a wetter climate for the middle Humboldt River based on high runoff and sediment transport, extensive stream migration and deposition of high-relief levees (Miller et al., 2004). The period from ~2600 to 2100 was one of the most intense periods of aggradation in the Middle Humboldt River during the late Holocene. A look at the geographic extent of the Humboldt River watershed (Figs. 1 and 8) reveals that it spans a region extending from ~39° to 42° N latitude. The northern headwaters of the Humboldt River are between 41 and 42° N latitude, towards the northern edge of the dipole transition zone. Paleocologic reconstructions from this latitude (e.g. Diamond Pond and Mission Cross Bog) show evidence for wet climate between ~2900 and 2000 cal yr BP, consistent with the findings of Miller et al. (2004) for extensive flooding of the main branch of the Humboldt River. In contrast, the headwaters of the Reese River tributary are in the Toiyabe Range, south of 40° N latitude and within the region that shows extensive evidence for drought. The river spans the dipole transition zone and illustrates why Miller et al. (2004) found the southern portion of the watershed to be in drought while the northern portion was in flood.

Along the Colorado and Salt River systems in Arizona, major floods were completely absent between 3900 and 2200 cal yr BP (Ely et al., 1993). Modern floods of this type are associated with a negative SOI and strong El Niño conditions. They argued that long-term flood variability is associated with persistent changes in regional climate and large scale atmospheric patterns. Thus, the period of no floods may have been dominated by a positive SOI and persistent La Niña type conditions. A persistent negative PDO pattern was suggested by Kirby et al. (2010) to explain low lake levels with no flood events in Lake Elsinore, southern California between ~3200 and 1960 cal yr BP. Low lake levels were also recorded in nearby Lower Bear Lake during this time period (Kirby et al., 2012).

We note that most of the sites in the southern Great Basin appear to have had an increase in effective moisture beginning ~2300 to 2200 cal yr BP and that by 2000 cal yr BP at Diamond Pond in the northern Great Basin had shifted to a drier climate (Wigand, 1987). This may indicate that by this time the pattern switched and the dipole had reversed. The Stonehouse Meadow record also shows a transitional period (Zone 3b) beginning ~2200 cal yr BP. It is still possible that drier than normal conditions continued for a while across the central Great Basin following a switch in the dipole because sites near the transition zone between 40 and 42° N (Fig. 7) can also receive anomalously low winter precipitation when under a negative SOI pattern (Wise, 2010).

The persistent regional pattern of dry and wet climate during the late Holocene suggests a long-term shift in atmospheric circulation across the region. It is likely that such persistent patterns are associated with changes within the oceans, however the relationships between ENSO, PDO, AMO, and dominant modes of atmospheric circulation patterns are not completely understood (McCabe et al., 2008). More Holocene records of ocean sea surface temperatures are needed to resolve this.

## 6. Conclusions

The Stonehouse Meadow record contains sedimentary, mollusk, diatom and pollen evidence for an extended dry period persisting from ~2800 to 1850 cal yr BP. Comparison of this record with multiple proxy records across the central Great Basin from the eastern Sierra Nevada to the Bonneville Basin suggests that this dry period was regional in extent. The beginning and ending dates vary among sites, but all sites record multiple centuries of dry climate between 2500 and 1900 cal yr BP. This duration makes it the longest persistent dry period within the late Holocene and suggests

long-term decreased average winter precipitation. In contrast, sites within the northern Great Basin experienced a wetter climate during this period.

The pattern of persistent wet and dry paleoclimate reconstructions is similar to the dipole pattern of ENSO-related precipitation variability across the west, with a dry southwest and wet northwest. The transition zone between wet and dry zones lies at approximately 40–42° N latitude; sites south of 40° N had persistent dry climate while those north of this transition had wetter climate. The wet/dry pattern is consistent with a weak Aleutian low shifted westward, the North Pacific high shifted north and decreased fall–winter SST off the California coast. The reconstructed pattern of precipitation in the Great Basin appears consistent with patterns associated with SST, high and low pressure systems, ENSO, and La Niña-type conditions with a northerly storm track, and an anomalously wet northwest and dry southwest.

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