A Reassessment of U-Th and \(^{14}\text{C}\) Ages for Late-Glacial High-Frequency Hydrological Events at Searles Lake, California

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INTRODUCTION

High-frequency climate fluctuations during the last glacial period, documented in Greenland ice core records and North Atlantic deep-sea sediments, have raised a keen interest among the paleoclimate community (Dansgaard et al., 1984; Oeschger et al., 1984; Bond et al., 1992; Johnsen et al., 1992; Chappellaz et al., 1993; Broecker, 1994). These abrupt changes present a new challenge to the understanding of the dynamics of our climate system, because they are not accounted for by the well-established orbital climate theory. Furthermore, the abrupt character of these changes bodes a possible societal threat as the anthropogenic greenhouse buildup continues. Increasing evidence has documented that these abrupt climatic changes were not restricted to the northern Atlantic region. Rather, their impacts appear to be worldwide (Chappellaz et al., 1993; Seltzer, 1994; Denton and Hendy, 1994; Gasse and Campo, 1994; Grimm et al., 1994; Broecker, 1994; Phillips et al., 1994; Clark and Bartlein, 1995; Lowell et al., 1995; Porter and An, 1995; Behl and Kennett, 1996; Benson et al., 1996; Philips et al., 1996). Although changes in the operation of thermohaline circula-
tion in the North Atlantic Ocean have been invoked as the cause of these fluctuations, there remains a missing mechanism linking the ocean changes to those in the atmosphere (Broecker, 1994). Broadly, the strategy to solve this mystery is to establish the spatial distribution of each event which may eventually allow links to be made. The focus of this study is the careful assessment of the chronology based on U-Th and \(^{14}\text{C}\) measurements for the hydrological events recorded in the Great Basin of the western United States.

In this paper we discuss the timing of the last, and most dramatic, pluvial event recorded in Lake Lahontan and Searles Lake, and of distinct dry periods recorded as evaporite layers in the Searles basin. The last major pluvial event in these closed basins happened during the transition interval between the last iceberg discharge event Heinrich 1 (~14,500 \(^{14}\text{C}\) yr; Bond et al., 1992; Bond and Lotti, 1995) and the Bolling-Allerød warming of ~12,800 \(^{14}\text{C}\) yr B.P (Siegenthaler et al., 1984).

The annual precipitation in the western Great Basin ranges from 20 to 40 cm and is mainly associated with winter westerly flow (Houghton et al., 1975; Thompson et al., 1993). In contrast, paleolake levels indicate that many closed-basin lakes in the Great Basin received enormous excess moisture during the last glacial period (Smith and Street-Perrott, 1983; Benson and Thompson, 1987). A climate with southward-displaced jet stream due to the presence of a larger Laurentide ice sheet has been proposed to account for this excess moisture (Hostetler and Benson, 1990; Thompson et al., 1993). Today the western Great Basin receives excess winter precipitation during the years of extreme El Niño Southern Oscillation (ENSO) (Ropelewski and Halpert, 1986). Could a climate pattern similar to that of the ENSO also have been associated with each pluvial event?

**SAMPLE LOCALITIES**

The last major pluvial event in Lake Lahontan left behind distinct high shorelines at about 1330 m (Benson et al., 1990). Tufa samples (impure calcium carbonate deposits) from these high shorelines in Lake Lahontan and from the shoreline at the Searles Lake spill-over level (~695 m) were collected for age determination (Figs. 1 and 2). Sediments thought to have been deposited during the last deep-lake phase exposed in Poison Canyon on the western edge of Searles Lake where the overflow of water from China Lake entered (Figs. 2 and 3) were sampled and analyzed for \(^{14}\text{C}\) and U-Th. This sequence is composed of mainly carbonate and opal-rich marls.

Lower Salt evaporite samples (Fig. 3) were obtained from a subsurface core (X-52) preserved at the North American Chemical Company in Trona, California. Radiocarbon ages of these salt layers from the same core were determined by Stuiver and Smith (1979). U-Th isotopic ratios measured by mass spectrometry were successfully obtained for six salt layers of the Lower Salt Unit.

**MATERIALS AND METHODS**

All new radiocarbon measurements reported here were carried out at the AMS facility at ETH in Switzerland. About 1–2 mg of carbon is needed for the AMS facility at ETH; hence, about 20 mg of each bulk calcium carbonate sample were processed. Acid leach experiments on massive carbonates were conducted in an attempt to reduce the secondary carbonate contamination (see Notes column in Table 1). Bulk marl samples were not leached before the conversion to \(\text{CO}_2\). Organic carbon in tufa is obtained from the acid insoluble residue of the bulk sample.

For U-Th isotopic measurements by mass spectrometry, about 0.5 g of each sample, spiked with a \(^{233}\text{U}-^{229}\text{Th}\) mixture prior to sample digestion, was totally dissolved in a mixture of \(\text{HNO}_3\), \(\text{HClO}_4\), and \(\text{HF}\). Extreme care was taken to ensure adding just enough HF to avoid precipitation of calcium.
fluoride. About 9 mg of FeCl₃ was added to precipitate the actinides by adding ammonium hydroxide. Chemical procedures for separating U and Th were the same as those used in Lin et al. (1996). Briefly, sample solution in 7N HNO₃ was loaded on 400 μl and then 100 μl anion exchange columns (AG 1 × 8 resin). Thorium was eluted with 6N HCl and U with 1 N HBr. Graphite colloids and phosphoric acid were added to the final solutions of Th and U, respectively, before mass spectrometry measurement. The blanks of U and Th were both about 10⁻² ng, negligible compared to the 10⁵ ng level of both U and Th of the samples.

Uranium and thorium isotopes were measured during two separate periods on two mass spectrometers. Before June 1994, the isotopic measurements were done on a VG Isolab 54 sector, where U isotopes were measured by thermal ionization (TIMS) and Th isotopes by secondary ionization (SIMS) techniques. The samples measured during this period include all the highstand tufa samples from Lake Lahontan and two Searles Lake samples (indicated in the footnotes of Table 2). Uranium and thorium of four Lower Salt evaporite samples (Table 2) were measured by TIMS in August 1995 on a VG54-30 sector (Appendix A). A more detailed description of VG Isolab 54 and internal calibrations for measuring Th and U isotopes can be found in Bourdon (1994) and in the Appendix A of Lin et al. (1996).

Additional U and Th analyses were carried out by alpha spectrometry; about 10 g of each carbonate sample was totally dissolved after adding ²³⁶U and ²²⁹Th. The analytical procedures for separating and purifying U and Th are described in Lao (1991). Purified U and Th were then electroplated onto silver disks and counted using Ortec or Tennelc silicon surface barrier alpha detectors. The blanks for U and Th were not detectable above counting background.

**RESULTS**

Radiocarbon results are listed in Table 1. U-Th isotopic results of tufas from the highest shorelines of Lake Lahontan were published previously (Lin et al., 1996). U-Th results of Searles Lake samples are shown in Table 2. The isochron of Lake Lahontan high shoreline samples yields a ²³⁰Th/²³⁴U age of 16,400 ± 700 (1σ) yr B.P.

The high shoreline tufa and Lower Salt samples from Searles Lake all have relatively low Th/U ratios (Table 2), and an initial ²³⁰Th correction was made by using an assumed initial ²³⁰Th/²³²Th ratio.
II. Searles Lake outcrop samples

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SL90-12 (bulk)

SL90-17-1 Terrace Hill 1326–1328 tufa 12,300 ± 100

SL90-17-2 Terrace Hill 1326–1328 tufa 12,200 ± 100

SL90-17-3 Terrace Hill 1326–1328 tufa 13,000 ± 100

SL90-18-2 Terrace Hill 1326–1328 tufa 12,700 ± 100

SL90-18-3 Terrace Hill 1326–1328 tufa 12,800 ± 100 4/5 leached

SL90-18-4 Terrace Hill 1326–1328 tufa 12,800 ± 100

SL90-18-5 Terrace Hill 1326–1328 tufa 12,300 ± 100

PL94-1 Terrace Hill 1328 o.m.2 in tufa 12,900 ± 120 HF leached

DISCUSSION

Reservoir Corrections for 14C Ages

The remnant lakes in the Lahontan and Owens River drainage systems (Searles Lake) are known to have lower 14C/12C ratios compared to those in the atmosphere (Broecker and Walton, 1959; Peng et al., 1978). Radiocarbon ages must therefore be corrected for this reservoir effect. Our only knowledge of the magnitude of the radiocarbon concentrations in these lakes comes from measurements of samples

TABLE 1

Pre-Reservoir-Corrected Radiocarbon Ages of Samples Associated with Lake Lahontan and Searles Lake Highstands

<table>
<thead>
<tr>
<th>Sample name</th>
<th>Locality</th>
<th>Altitude (m)</th>
<th>Material</th>
<th>14C age (yr B.P.)</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>I. Lake Lahontan</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Pyramid Lake high shorelines</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>JL90-5</td>
<td>Marble Bluff</td>
<td>1330</td>
<td>tufa</td>
<td>13,100 ± 100</td>
<td></td>
</tr>
<tr>
<td>JL90-6</td>
<td>Terrace Hill</td>
<td>1330</td>
<td>tufa</td>
<td>13,200 ± 100</td>
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</tr>
<tr>
<td>JL90-15</td>
<td>Terrace Hill</td>
<td>1330</td>
<td>tufa</td>
<td>12,200 ± 100</td>
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<tr>
<td>JL90-16</td>
<td>Terrace Hill</td>
<td>1330</td>
<td>tufa</td>
<td>12,900 ± 100</td>
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</tr>
<tr>
<td>JL90-17-1</td>
<td>Terrace Hill</td>
<td>1326–1328</td>
<td>tufa</td>
<td>12,300 ± 100</td>
<td></td>
</tr>
<tr>
<td>JL90-17-2</td>
<td>Terrace Hill</td>
<td>1326–1328</td>
<td>tufa</td>
<td>12,200 ± 100</td>
<td></td>
</tr>
<tr>
<td>JL90-17-3</td>
<td>Terrace Hill</td>
<td>1326–1328</td>
<td>tufa</td>
<td>13,000 ± 100</td>
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</tr>
<tr>
<td>JL90-18-2</td>
<td>Terrace Hill</td>
<td>1326–1328</td>
<td>tufa</td>
<td>12,700 ± 100</td>
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<tr>
<td>JL90-18-3</td>
<td>Terrace Hill</td>
<td>1326–1328</td>
<td>tufa</td>
<td>12,800 ± 100</td>
<td>4/5 leached</td>
</tr>
<tr>
<td>JL90-18-4</td>
<td>Terrace Hill</td>
<td>1326–1328</td>
<td>tufa</td>
<td>12,800 ± 100</td>
<td></td>
</tr>
<tr>
<td>JL90-18-5</td>
<td>Terrace Hill</td>
<td>1326–1328</td>
<td>tufa</td>
<td>12,300 ± 100</td>
<td></td>
</tr>
<tr>
<td>PL94-1</td>
<td>Terrace Hill</td>
<td>1328</td>
<td>o.m.2 in tufa</td>
<td>12,900 ± 120</td>
<td>HF leached</td>
</tr>
<tr>
<td>2. Carson Sink high shorelines</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CSLB(U) Jessup</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SL93-12</td>
<td>Navy Road</td>
<td>~695</td>
<td>tufa</td>
<td>12,400 ± 100</td>
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</tr>
<tr>
<td>C3c</td>
<td>Poison Canyon</td>
<td>~590</td>
<td>bulk marl</td>
<td>11,200 ± 100</td>
<td></td>
</tr>
<tr>
<td>C3b</td>
<td>Poison Canyon</td>
<td>~590</td>
<td>bulk marl</td>
<td>12,200 ± 100</td>
<td></td>
</tr>
<tr>
<td>C3a</td>
<td>Poison Canyon</td>
<td>~590</td>
<td>bulk marl</td>
<td>12,300 ± 100</td>
<td></td>
</tr>
<tr>
<td>C2</td>
<td>Poison Canyon</td>
<td>~587</td>
<td>shell</td>
<td>12,100 ± 100</td>
<td>1/3 leached</td>
</tr>
<tr>
<td>C2</td>
<td>Poison Canyon</td>
<td>~587</td>
<td>shell, sand</td>
<td>12,300 ± 100</td>
<td>1/5 leached</td>
</tr>
<tr>
<td>C1c</td>
<td>Poison Canyon</td>
<td>~585</td>
<td>bulk marl</td>
<td>13,200 ± 100</td>
<td></td>
</tr>
<tr>
<td>C1a</td>
<td>Poison Canyon</td>
<td>~585</td>
<td>bulk marl</td>
<td>12,900 ± 100</td>
<td></td>
</tr>
<tr>
<td>SL93-21 (base of C1)</td>
<td>Poison Canyon</td>
<td>~579</td>
<td>bulk marl</td>
<td>13,800 ± 95</td>
<td></td>
</tr>
<tr>
<td>Bc</td>
<td>Poison Canyon</td>
<td>~575</td>
<td>bulk marl</td>
<td>16,090 ± 130</td>
<td></td>
</tr>
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</table>

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TABLE 3
The Reservoir Ages of Prebomb Pyramid Lake and Estimated Values for Lake Lahontan High Shoreline Samples

<table>
<thead>
<tr>
<th></th>
<th>(G^*) (mol/m²/yr)</th>
<th>(R) (mol/m²/yr)</th>
<th>(\Delta^{14}C) (%)</th>
<th>(C_L/C_A)</th>
<th>Reservoir age (yr B.P.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>I. Modern Pyramid Lake (prebomb)</td>
<td></td>
<td></td>
<td>-80</td>
<td>0.93</td>
<td>590</td>
</tr>
<tr>
<td>Measured</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>Estimate 1'</td>
<td>0</td>
<td>8</td>
<td>—</td>
<td>0.98</td>
<td>180</td>
</tr>
<tr>
<td>Estimate 2'</td>
<td>0.48</td>
<td>8</td>
<td>—</td>
<td>0.93</td>
<td>590</td>
</tr>
<tr>
<td>II. Lake Lahontan last highstand at 1330 m</td>
<td></td>
<td></td>
<td>—</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>Estimate 1'</td>
<td>0</td>
<td>8</td>
<td>—</td>
<td>0.97</td>
<td>260</td>
</tr>
<tr>
<td>Estimate 2'</td>
<td>0.48</td>
<td>8</td>
<td>—</td>
<td>0.92</td>
<td>680</td>
</tr>
<tr>
<td>Estimate 3'</td>
<td>0</td>
<td>17</td>
<td>—</td>
<td>0.98</td>
<td>162</td>
</tr>
</tbody>
</table>

* \(G^*\) is the hypothesized \(^{14}\)C-free flux of CO\(_2\) from hot springs.

\[ \frac{C_L}{C_A} = \frac{R + l_E k_R C_R}{R + l_E k_R + G + \frac{V_L}{A_L} k_L}, \]

where \(C_L\) and \(C_A\) are \(^{14}\)C concentrations of the \(\Sigma\)CO\(_2\) in the lake water (Broecker and Walton, 1959) and atmosphere. \(R\) is the gas exchange rate of CO\(_2\). \(l_E\), \(k_R\) and \(C_R\) are the linear evaporation rate (m/yr) from the lake surface, concentration of \(\Sigma\)CO\(_2\) in the incoming river and \(^{14}\)C concentration of the river. \(V_L\) and \(k_L\) are volume of the lake and \(\Sigma\)CO\(_2\) concentration in the lake. \(\lambda\) is the decay constant of \(^{14}\)C.

\(^a\) Assumed an atmospheric \(\Delta^{14}C\) of \(-10\)%.

\(^b\) Estimated values are calculated from Eq. (T1), with modern linear evaporation of \(l_E = 1.3\) m/yr and a \(V_L/A_L = 38\) m. \(C_L/C_A\) is the measured value of 0.92 (Broecker and Walton, 1959).

\(^c\) Estimated values are calculated from Eq. (T1), with modern linear evaporation of \(l_E = 0.75\) m/yr and a \(V_L/A_L = 91\) m (Benson, 1993). The rest of the parameters in Eq. (1) for this estimate are also adopted from Benson (1993).

\(^*\) This calculation uses a jet stream \(l_E\) of 0.75 m/yr and a \(V_L/A_L\) of 91 m (Benson, 1993). The rest of the parameters in Eq. (1) for this estimate are also adopted from Benson (1993).

Modern Pyramid Lake serves as an analog for Lake Lahontan, while neighboring Mono Lake has a \(\Sigma\)CO\(_2\) similar to that reconstructed for Searles Lake before its last desiccation (Smith, 1979; Peng et al., 1978). Radiocarbon budgets for modern Mono Lake and Pyramid Lake appear to require significant additions of radiocarbon-free CO\(_2\) from springs. In the case of Mono Lake, a radiocarbon-free carbon flux of 0.9 mol/m²/yr was needed to explain the radiocarbon measurements for the prebomb lake water with a gas exchange rate of 7 mol/m²/yr obtained from an SF\(_6\) tracer experiment (Wanninkhof, 1987). Model results suggest that the p CO\(_2\) of Mono Lake water prior to 1945 (when the source water was diverted into the Los Angeles Aqueduct) was 1.3 times that of the atmosphere and requires an input of CO\(_2\) of about 3.3 mol/m²/yr (Oxburgh et al., 1991). For Pyramid Lake, a flux of radiocarbon-free CO\(_2\) of 0.48 mol/m²/yr is needed to account for the measured prebomb \(^{14}\)C concentration of the total CO\(_2\) in lake water (Table 3). A gas exchange rate (\(R\)) of 8 mol/m²/yr obtained from new results of SF\(_6\) experiment in Pyramid Lake is adopted in the calculation of reservoir age using Eq. (T1) in Table 3 (R. Wanninkhof, personal communication, 1995). Without the \(^{14}\)C-free CO\(_2\) flux from springs (\(G^*\) term in Eq. (T1)), the calculated reservoir age is only 180 yr, while the observed value was 590 yr (Broecker and Walton, 1959).

In the estimate of paleoreservoir age for Lake Lahontan, two cases were calculated, one with a \(^{14}\)C-free CO\(_2\) flux (\(G\) term) of 0.48 mol/m²/yr and one without a \(G\) term. This first yields a reservoir age of 680 yr for the high shoreline samples, compared to the 260 yr obtained without the input of the \(G\) term (Table 3). The first case refers to a situation where the input of \(^{14}\)C-free CO\(_2\) from spring water increased linearly with the expanded lake area, which is assumed to be a maximum of this flux. We adopted the mean value between the reservoir ages calculated with the maximum \(^{14}\)C-free CO\(_2\) flux and without this flux, so the reservoir age used to correct all \(^{14}\)C ages becomes 470 ± 200 yr. Benson (1993) used a higher CO\(_2\) exchange rate, ignoring the radiocarbon-free CO\(_2\) input, and obtained a reservoir correction of ~160 yr.

Although Searles Lake had a large \(\Sigma\)CO\(_2\) inventory (Smith, 1979), the evidence seems to suggest no additional input of carbon from sources other than the Owens River (Smith, 1976). Thus, the estimate of reservoir correction for Searles Lake is carried out as established by Peng et al. (1978), where gas exchange rate and total CO\(_2\) in the lake are the
two main factors controlling the exchange of $^{14}$C between the lake water and the atmosphere. The reservoir ages for C unit samples calculated this way are shown in Table 4 for different lake surface areas. Using Eq. (T2) in Table 4, the $N$ term (ΣCO$_2$) in the calculation of reservoir age for C unit samples (Parting Mud) was estimated from the total carbonate ions (CO$_2$$^-$ and HCO$_3^-$) in the Upper Salt Unit, which is a unit of massive evaporites deposited right after the C unit (2.7 × 10$^{12}$ mol Smith, 1979).

If the pluvial event that deposited the C unit reached the spill level of Searles Lake when the lake surface area was 1000 km$^2$ (Smith and Street-Perrott, 1983), then the reservoir age is about 330 yr (gas exchange rate of 8 mol/m$^2$/yr; Table 4). The same calculation for Lower Salt samples with different lake surface areas. Using Eq. (T2) in Table 4, the reservoir age is derived from the total carbonate ions (CO$_2$$^-$ and HCO$_3^-$) in the Upper Salt Unit (2.7 × 10$^{12}$ mol Smith, 1979).

Upper Salt unit samples calculated this way are shown in Table 4 for different lake surface areas. Using Eq. (T2) in Table 4, the reservoir age is derived from the total carbonate ions (CO$_2$$^-$ and HCO$_3^-$) in the Upper Salt Unit (2.7 × 10$^{12}$ mol Smith, 1979).

$\Lambda$ is the invasion rate of CO$_2$ into the lake and is equal to $RA_l$. (Peng et al., 1978). $N$ is the amount of total CO$_2$ in the lake in mol/m$^2$. The amount of total CO$_2$ in the lake at the time of the deposition of C unit is estimated from the total carbonate ions (CO$_2$$^-$ and HCO$_3^-$) in the Upper Salt Unit, which is a unit of massive evaporites deposited right after the C unit (2.7 × 10$^{12}$ mol Smith, 1979).

$\alpha$ is the amount of total CO$_2$ in the lake in mol/m$^2$. The lake would have been enriched in dissolved U, thus resulting in elevated $^{230}$Th/$^{232}$Th of the dissolved Th in the lake (Lin et al., 1996). However, the Ri values might have changed during the interval of Lower Salt deposition, since the fluctuating hydrological balances may have caused the lake water chemistry to vary, especially in the content of carbonate ions, which in turn affected the solubilities of U and Th. We thus adopt a range of Ri values for each sample until more precise estimates of the Ri's in Searles Lake can be obtained.

**230$^{\text{Th}}$ Age Determinations for Searles Lake Samples**

Corrected $^{230}$Th ages for Searles Lake tufa and salt samples depend on the initial $^{230}$Th/$^{232}$Th activity ratios (Fig. 4). Peng et al. (1978) used 1.7 as the initial $^{230}$Th/$^{232}$Th activity ratio, which was determined from Lake Lahontan carbonates. As discussed in Lin et al. (1996), the initial Th in the tufa samples from Lake Lahontan may contain a significant contribution from a hydrogenous phase due to the high dissolved Th content in the alkaline water. For the even more alkaline condition that may have existed in Searles Lake, the initial Th in the impure carbonates was probably also composed of both lithologic and hydrogenous components, perhaps toward a higher hydrogenous component than in Lake Lahontan (higher initial $^{230}$Th/$^{232}$Th). It is thus necessary to determine this value for the Searles Lake samples.

The initial $^{230}$Th/$^{232}$Th ratio of the Lower Salt samples is constrained by the requirement of maintaining stratigraphic order of all samples. It can be seen in Figure 4 that the $^{230}$Th ages for S4, S5, and S6 are less sensitive to the initial $^{230}$Th/$^{232}$Th activity ratios (Ri) applied, due to higher U/Th ratios of these samples than in others (Table 2). Two pairs of samples yield stratigraphically consistent ages only in certain ranges of Ri’s, i.e., Ri greater than 1.82 for S7 and S6, and Ri smaller than 1.84 for and S4 with S3 (Fig. 4). If a uniform initial $^{230}$Th/$^{232}$Th ratio among all samples were the case, the value of about 1.83 would be the best choice of Ri. This value is slightly higher than the average Ri of Lake Lahontan samples (1.6−1.7, Lin et al., 1996). This result is not surprising, as the supposedly high alkaline condition of Searles Lake would have been enriched in dissolved U, thus resulting in elevated $^{230}$Th/$^{232}$Th of the dissolved Th in the lake (Lin et al., 1996). However, the Ri values might have changed during the interval of Lower Salt deposition, since the fluctuating hydrological balances may have caused the lake water chemistry to vary, especially in the content of carbonate ions, which in turn affected the solubilities of U and Th. We thus adopt a range of Ri values for each sample until more precise estimates of the Ri’s in Searles Lake can be obtained.

**TABLE 4**

*Estimates of Reservoir Ages for C Unit Samples from Searles Lake, Using Total CO$_2$ in the Upper Salt Unit*

<table>
<thead>
<tr>
<th>Lake depth (m)</th>
<th>Lake surface area (km$^2$)</th>
<th>Reservoir age (yr) $\text{R} = 6$</th>
<th>Reservoir age (yr) $\text{R} = 7$</th>
<th>Reservoir age (yr) $\text{R} = 8$</th>
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<tbody>
<tr>
<td>170</td>
<td>600</td>
<td>714</td>
<td>616</td>
<td>542</td>
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<tr>
<td>175</td>
<td>700</td>
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<td>466</td>
</tr>
<tr>
<td>180</td>
<td>800</td>
<td>542</td>
<td>466</td>
<td>410</td>
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<td>190</td>
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<tr>
<td>200</td>
<td>1000</td>
<td>436</td>
<td>375</td>
<td>329</td>
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</table>

*Calculated from $\frac{C_l}{C_s} = \frac{1}{1 + \lambda N}$

$\lambda$ is the invasion rate of CO$_2$ into the lake and is equal to $RA_l$. (Peng et al., 1978). $N$ is the amount of total CO$_2$ in the lake in mol/m$^2$. The amount of total CO$_2$ in the lake at the time of the deposition of C unit is estimated from the total carbonate ions (CO$_2$$^-$ and HCO$_3^-$) in the Upper Salt Unit, which is a unit of massive evaporites deposited right after the C unit (2.7 × 10$^{12}$ mol Smith, 1979).

**FIG. 4.** Variation of corrected $^{230}$Th/$^{234}$U ages for the high shoreline tufa (SL93-12) and Lower Salt samples (S2 to S7) from Searles Lake as a function of the initial $^{230}$Th/$^{232}$Th activity ratios (Ri) applied, due to higher U/Th ratios of these samples than in others (Table 2). Two pairs of samples yield stratigraphically consistent ages only in certain ranges of Ri’s, i.e., Ri greater than 1.82 for S7 and S6, and Ri smaller than 1.84 for and S4 with S3 (Fig. 4). If a uniform initial $^{230}$Th/$^{232}$Th ratio among all samples were the case, the value of about 1.83 would be the best choice of Ri. This value is slightly higher than the average Ri of Lake Lahontan samples (1.6−1.7, Lin et al., 1996). This result is not surprising, as the supposedly high alkaline condition of Searles Lake would have been enriched in dissolved U, thus resulting in elevated $^{230}$Th/$^{232}$Th of the dissolved Th in the lake (Lin et al., 1996). However, the Ri values might have changed during the interval of Lower Salt deposition, since the fluctuating hydrological balances may have caused the lake water chemistry to vary, especially in the content of carbonate ions, which in turn affected the solubilities of U and Th. We thus adopt a range of Ri values for each sample until more precise estimates of the Ri’s in Searles Lake can be obtained.
The corrected $^{14}$C and $^{230}$Th ages listed in Table 5. The corrected $^{14}$C and $^{230}$Th ages, along with the corals results from Bard et al. (1993) are plotted on Figure 5 for comparison.

### Last Major Pluvial Event in Western Great Basin

The $^{230}$Th isochron age for Lake Lahontan high shoreline tufa samples is $16,400 \pm 700$ yr and the two corrected $^{230}$Th ages of the Searles Lake high shoreline tufa sample are $16,500 \pm 560$ and $17,000 \pm 1250$ yr B.P. (SL93-12 in Table 2). Garcia et al. (1993) obtained a $^{230}$Th age of $17,000 \pm 1000$ yr B.P. for three high shoreline tufa samples by alpha-counting technique, where no correction for initial $^{230}$Th was applied due to their high $^{230}$Th/$^{232}$Th ratios. After reservoir corrections ($470 \pm 210$ yr for Lake Lahontan samples and $560 \pm 17,000$ yr for Searles Lake samples), the two sets of independently obtained $^{14}$C,$^{230}$Th age pairs are consistent, with mean values falling below the coral trend by at least $1500$ $^{14}$C yr (Fig. 5). Although $^{230}$Th ages have relatively high uncertainties, the consistency among all $^{14}$C,$^{230}$Th age pairs points to

### Table 5

<table>
<thead>
<tr>
<th>Sample name</th>
<th>$^{14}$C age (yr B.P.)</th>
<th>Reservoir corr. $^{14}$C age (yr B.P.)</th>
<th>Corr. $^{230}$Th ages isochron (yr B.P.)</th>
<th>Corr. $^{230}$Th ages variable Ri’s (1–2.5)$^a$ (yr B.P.)</th>
<th>Corr. $^{230}$Th ages Peng et al., 1978$^b$ (yr B.P.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Lake Lahontan last highstand tufa</td>
<td>12,700 ± 350</td>
<td>12,200 ± 560</td>
<td>16,400 ± 700</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Note: The reservoir ages for Lake Lahontan high shoreline samples and Searles Lake’s C unit samples (including high shoreline tufa, equivalent to the upper most Parting Mud unit, Fig. 3) are $470 \pm 210$ and $330$ years, respectively. Reservoir age for Lower Salt evaporites is $700$ yr (see discussion in the text). All age uncertainties are $1\sigma$ errors.

$^a$ Age uncertainties shown are results of the range of Ri’s applied plus analytical errors.

$^b$ Data of samples from the same core as in this study, X-52; Ri = 1.7 was used for age correction.

$^c$ Interpolated from the $^{14}$C ages of adjacent mud layers (Stuiver and Smith, 1979; Peng et al., 1978).

independently. The range of 1–2.5 for the Ri values was chosen somewhat arbitrarily. The lower limit is a value between the average values of upper crust and igneous rocks (0.6–1) and detrital sediments in the Lahontan basin (1.17; Lin et al., 1996). An upper limit of 2.4 must be assigned for the Ri values for S3 and S4 are no greater than 1.84 (1–1.84), in order to maintain the stratigraphic order of all three samples (Fig. 4). For S5, a range from 1 to 2.5 was applied in the age correction, whereas for S6 and S7 the range of 1.82–2.5 was used. The $^{230}$Th ages thus corrected are listed in Table 5. The age uncertainties (1σ) listed in Table 3 include the effects of using a range of Ri values together with the analytical errors.

### $^{14}$C–$^{230}$Th Age Comparison and Timing of Hydrologic-Climatic Events

The $^{14}$C and $^{230}$Th ages of all samples obtained in this study are listed in Table 5. The corrected $^{14}$C and $^{230}$Th ages,
oldest $^{14}$C ages of the last highstand samples from both lakes suggest an age for the last pluvial event in both lakes of between 12,000 and 12,800 $^{14}$C yr B.P. after reservoir age correction (Tables 1 and 5), still younger than the $^{14}$C age converted from the $^{230}$Th age by 800 to 1600 $^{14}$C yr (Fig. 6). However, these comparisons contain several sources of uncertainties that may account for possible errors of $\sim 300$ yr for $^{14}$C ages (mainly from reservoir correction) and $\sim 600$ yr, at a minimum, for the U-Th chronology. On the other hand, Benson (1993) determined the timing of the last highstand of Lake Lahontan to be $\sim 13,800$ $^{14}$C yr B.P., based on $^{14}$C ages of tufa samples from the Walker Lake subbasin, which is consistent with the age suggested by $^{230}$Th age of highstand tufas but significantly older than the $^{14}$C ages obtained in this study (Fig. 6).

The $^{14}$C age of the last major pluvial event in Searles Lake is consistent with the intensive wetness observed in Owens Lake between 12,000 and 13,000 $^{14}$C yr B.P. (Benson et al., 1996). A more detailed correlation between the oxygen isotope records in Searles Lake and Owens Lake during this period will be discussed in a separate paper prepared by F. M. Phillips (personal communication, 1997).

The ‘‘Trans-US Wet’’ Event

While the major pluvial event was taking place in the Great Basin between 12,000 and 13,000 $^{14}$C yr ago, a large corresponding peak was not present in the amount of dust blown over Greenland (Mayewski et al., 1994). However, a massive iceberg discharge was launched into the north Atlantic just prior to 14,000 $^{14}$C yr ago (Heinrich 1 event, Bond...
Some climatic factors may have been in an unstable state during this interval, and were reflected in Laurentide ice sheet instability and the hydrological cycles on the North American continent.

Evidence is accumulating for an approximately synchronous climate anomaly in many parts of the world between 13,000 and 15,000 $^{14}$C yr ago. We have compiled a preliminary map showing the spatial pattern of this event in Figure 7 (upper panel). A noticeable wet event during this period extended across the southwestern and southeastern regions of the contiguous United States, Broecker (1994) has called this the “Trans-US wet event.” The global distribution of the modern ENSO-related climate anomalies during typical El Niño years is also shown in Figure 7 for comparison (lower panel). The pattern of this late-glacial climate event is particularly similar to the modern ENSO pattern in the North America and Australia, but not in the North Atlantic region (cold instead of no connection with modern ENSO) and equatorial South America (wet instead of dry) (Fig. 7). This comparison highlights the importance of reconstructing the atmospheric circulation pattern of the late-glacial abrupt climatic changes. Perhaps a process akin to the teleconnection mechanism associated with the modern ENSO events (Hamilton and Garcia, 1986; Lau and Nath, 1994) played a role in the late-glacial climate but in a somewhat different way. The work of documenting spatial distribution and timing of late-glacial millennial-scale climate events should be continued.

### Lower Salt Unit of Searles Lake and Correlation with Other Climate Records

The $^{14}$C ages and $^{230}$Th ages of units S7, S6, and S5 (Fig. 3) fall close to the concordia line within 2σ errors. For S4, S3, and S2, the mean $^{230}$Th ages exceed $^{14}$C ages, but with large age uncertainties (Fig. 5). If the $^{14}$C,$^{230}$Th calibration
FIG. 8. Correlations of the Santa Barbara Basin Site 893 bioturbation index, \( \delta^{18} \text{O} \) record of the Summit Greenland ice core, and Searles Lake mud–salt oscillations (Behl and Kennett, 1996; Johnsen et al., 1992). The lengths of the arrows representing dust peaks are approximately proportional to the magnitudes of the dust concentration (from Mayewski et al., 1994).

curve of coral data were extended to 40,000 cal yr B.P., the calibration curve would probably agree better with the \( ^{14} \text{C} \)-\( ^{230} \text{Th} \) pairs of S4, S3, and S2, with \( ^{230} \text{Th} \) ages exceeding \( ^{14} \text{C} \) ages by 2000 to 3000 yr (Fig. 5). However, many factors could have altered the \( ^{14} \text{C} \) ages of Lower Salt samples, including contamination from old as well as young secondary carbon (Stuiver and Smith, 1978) and an inaccurate reservoir correction, plus the relatively large uncertainties for some \( ^{230} \text{Th} \) ages. Accordingly, we will reserve final judgment on the comparison between \( ^{230} \text{Th} \) and \( ^{14} \text{C} \) ages.

The alternating wet–dry conditions of Searles Lake represented by the Lower Salt unit bears a strong resemblance to the abrupt climate changes recorded by the Greenland ice core records (Dansgaard et al., 1984; Johnsen et al., 1992; Phillips et al., 1994), North Atlantic deep sea sediments (Bond et al., 1992), and the fluctuations in oxidation conditions in the Santa Barbara Basin sediments caused by rapid circulation changes in the basin (Behl and Kennett, 1996) (Fig. 8). Although the new \( ^{230} \text{Th} \) ages cannot resolve the timing of the dry events in Searles Lake to better than 1000 yr, due to the uncertainty in the initial \( ^{230} \text{Th}/^{232} \text{Th} \) of Searles Lake samples, the age ranges constrained here do offer a provocative correlation with the Greenland ice-core records.

Phillips et al. (1994) proposed a link between the Searles Lake hydrological balance and the \( \delta^{18} \text{O} \) record of Greenland Summit ice cores using the \( ^{230} \text{Th} \) alpha counting ages of the Lower Salt samples obtained by Peng et al. (1978). The \( ^{230} \text{Th} \) age ranges of Lower Salt samples obtained in this study suggest a different correlation than the one proposed by Phillips et al. (1994). We first assume that, like the Holocene, late Pleistocene dry conditions in the Great Basin lakes generally corresponded to warm phases in the North Atlantic basin. For instance, the Parting Mud Unit, representing a period of relatively wet climate in Searles Lake, corresponded to the last glacial maximum and the several thousand yr following. Because units S7 and S6 are constrained to be no older than 24,400 cal yr B.P. (Table 5), they appear to correlate with interstad 2 (double peaks between 23,000 and 23,500 cal yr B.P., Fig. 8). The age range of S5 covers the whole interval of interstad es 3 and 4, with the mean age corresponding to interstad e 3. Units S4 and S3 may correlate with interstad es 5 and 6. The age range of S2 corresponds to interstad es 7 and 8. Overall, the age ranges for S6 and S4 are considered to be the best constrained, with the corresponding ice-core events being interstad es 2 and 5 (Fig. 8).

Below unit S1 of the Lower Salt, the stratigraphy of
Searles Lake does not show distinct wet–dry episodes like the Lower Salt unit. This may be partially due to the generally deeper lake conditions while the Bottom Mud unit was deposited (Smith, 1979). In other words, the distinctly interbedded mud–salt layers could not have happened in a drainage basin with a continuously high water balance, whereas Searles Lake, being the terminal lake of the drainage system at the time, had the highest sensitivity to any changes of the climate system that affected its water balance. The apparent linkage between Searles Lake and the Greenland ice-core records leads to a correlation between the Searles Lake hydrological balance and the dust record in Greenland ice cores. It is not hard to comprehend this linkage when the similar mechanisms behind the two records are considered. Like the dust content in Greenland ice cores, the Great Basin hydrological balance is also associated with westerly winds (Mayewski et al., 1994; Thompson et al., 1993). If the presumption that the wet phases in Searles Lake generally correlate with the stadial events in the ice-core records is valid, then the dust peaks are thus correlated with the wet phases of the Great Basin lakes. This relationship between ice-core dust content and Searles Lake hydrology appears to hold throughout the late-glacial period and the Holocene, with a nonlinear correlation (Fig. 8; data from Smith and Street-Perrott, 1983; Mayewski et al., 1994). For example, the times of most prominent dust peaks (i.e., ca. 25,000 and 24,000 cal yr B.P.) corresponded to low or intermediate lake levels in Searles Lake and Lake Lahontan (Smith and Street-Perrott, 1983; Benson et al., 1990). What we learn from the above observation in terms of the worldwide atmospheric circulation during this period are (1) the mid-latitude westerlies seem to be playing an active role during the late-glacial interval of abruptly changing climate, and (2) the different components in the westerlies that controlled the amounts of dust transported into Greenland and moisture into the Great Basin did not always correspond to the climate system with proportional magnitudes.

**CONCLUSIONS**

Closed-basin lakes in the western Great Basin, as in many other regions of the world, experienced prominent millennial-scale climate fluctuations during late-glacial time. The U-Th isochron ages for shoreline tufas from both Lake Lahontan and Searles Lake suggest that the last major pluvial event in these two basins was synchronous at ca. 16,500 cal yr B.P., equivalent to a radiocarbon age of 14,000–13,500 yr B.P. Whereas radiocarbon ages determined for samples recording the last highstand of both lakes suggest that this pluvial event occurred between 12,800 and 12,000 yr B.P. Disagreement with the $^{14}$C-$^{230}$Th ages curve of Barbados corals is on the order of 1000 $^{14}$C yr.

Distinct evaporite layers in an interbedded mud–salt unit of Searles Lake were deposited as a result of periodic near-drying conditions in Searles Lake. U-Th dating of these salt layers reveals that millennial-scale wet–dry events in Searles Lake occurred between about 24,000 and 35,000 yr B.P. and correlate approximately with the Dansgaard-Oeschger cycles between interstades 2 and 8 in the Greenland Summit ice cores.

**APPENDIX**

**Mass Spectrometry Measurements of U-Th Isotopes**

For the samples measured by Secondary Ionization Mass Spectrometry on VG Isolab 54, the final solution of Th is loaded onto graphite rods to minimize isobaric interference (Secondary Ionization Mass Spectrometry) with an Ar$^+$ beam (primary beam). Uranium isotopes are loaded with graphite colloid on a single Re and run thermally. A multi-collection analysis on Faraday and Daly collectors is used to measure both U and Th isotopes, where the high abundance isotopes ($^{230}$Th and $^{238}$U) were collected on Faraday cups and the low abundance isotopes ($^{229}$Th, $^{234}$U, $^{233}$U, $^{235}$U, and $^{233}$U) were collected on Daly detector simultaneously. The Faraday-Daly gain was measured to within 0.5% before starting sample measurement and measured repeatedly before and after each run. An Electrostatic Analyser (ESA) before the Daly detector reduces the tailing ions of a high abundance isotope at the nearby mass of a low abundance isotope. The abundance sensitivity is about 0.1 ppm at 2 amu. The external reproducibility of the atomic $^{230}$Th/$^{232}$Th and $^{234}$U/$^{238}$U ratios, both on the magnitude of $10^{-5}$, were 0.5% or better at 2σ uncertainty level. Standards for U and Th were measured frequently to ensure the proper performance of the machine. The standard values are $5.43 \pm 0.03 \times 10^{-5}$ for $^{234}$U/$^{238}$U (LaU03) and $6.42 \pm 0.03 \times 10^{-6}$ for $^{230}$Th/$^{232}$Th (LaTh1).

Uranium and Thorium of four Lower Salt evaporite samples were measured thermally in August 1995 on VG54-30 at Lamont. The machine is also equipped with an ESA filter before the Daly detector for high abundance sensitivity measurements. Both U and Th were loaded with graphite colloid on single Re filament. All U and Th isotopes were collected on Daly detector with peak switching routines. An U ion beam of about tenths of a mV to several mV intensities could be acquired with the filaments heated to 4.4 to 4.5 amps. Ion beam of Th is usually from several mV to several tens of mV with currents of about 5 amps. Mass fractionation of U isotopes were not corrected during the isotopic measurement, but was corrected by the isotopic measurements of an U standard (NBS U500) using the same equation shown in Appendix A of Lin et al. (1996). The correction was about 0.16% per amu. The mass fractionation correction for Th isotopes was not done, but is considered insignificant compared to the external reproducibility. Data acquisition time for both U and Th was about 2 to 3 h. External reproducibilities of both U
and Th isotopic ratios were about 0.7% (2σ) and have been incorporated into the results reported in Table 2.

ACKNOWLEDGMENTS

Dr. F. Phillips is highly credited for initiating the idea of reapproaching and the age problems of Lower Salt samples with high precision dating technique, and we thank him for helping with sample collections in the field during the 1993 and 1994 expeditions. We thank G. Moulton of the North American Chemical Company for providing evaporite samples from the Lower Salt unit and Dr. L. Benson and Dr. S. Wisnouski for helping with the Lake Lahontan high shoreline tufa sampling. The first author also appreciates J. Rubenstone, B. Bourdon, A. Zindler, M. Fleisher, J. Clark, M. Class, T. Liu, D. Peteet, and M. Kneller for helpful technical suggestions and/or intriguing scientific insights about the paper. We appreciate the thorough and constructive reviews from Dr. G. I. Smith, Dr. S. Porter, and an anonymous referee. This project was mainly supported by NASA Grant No. NCC 5-29c at Lamont-Doherty Earth Observatory.

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