Journal of Paleolimnology 31: 275–293, 2004.



# Late Holocene lake level dynamics inferred from magnetic susceptibility and stable oxygen isotope data: Lake Elsinore, southern California (USA)

Matthew E. Kirby[1,*], Christopher J. Poulsen[2], Steve P. Lund[3], William P. Patterson[4], Liam Reidy[5] and Douglas E. Hammond[3]

[1]*Department of Geological Sciences, California State University, Fullerton, Fullerton, CA 92834, USA;* [2]*Department of Geological Sciences, University of Michigan, Ann Arbor, MI 48109, USA;* [3]*Department of Earth Sciences, University of Southern California, Los Angeles, CA 90089, USA;* [4]*Department of Geological Sciences, University of Saskatchewan, Saskatoon, SK, Canada S7N 5E2;* [5]*Department of Geography, University of California, Berkeley, 507 McCone Hall, Berkeley, CA 94720, USA;* **Author for correspondence (e-mail: mkirby@fullerton.edu)*

Received 25 March 2003; accepted 3 July 2003

*Key words:* Lake Elsinore, Lake level, Isotopes, Magnetic susceptibility, Precipitation, Southern California

**Abstract**

Southern California faces an imminent freshwater shortage. To better assess the future impact of this water crisis, it is essential that we develop continental archives of past hydrological variability. Using four sediment cores from Lake Elsinore in Southern California, we reconstruct late Holocene (~3800 calendar years B.P.) hydrological change using a twentieth-century calibrated, proxy methodology. We compared magnetic susceptibility from Lake Elsinore deep basin sediments, lake level from Lake Elsinore, and regional winter precipitation data over the twentieth century to calibrate the late Holocene lake sediment record. The comparison revealed a strong positive, first-order relationship between the three variables. As a working hypothesis, we suggest that periods of greater precipitation produce higher lake levels. Greater precipitation also increases the supply of detritus (i.e., magnetic-rich minerals) from the lake's surrounding drainage basin into the lake environment. As a result, magnetic susceptibility values increase during periods of high lake level. We apply this modern calibration to late Holocene sediments from the lake's littoral zone. As an independent verification of this hypothesis, we analyzed $\delta^{18}O_{(calcite)}$, interpreted as a proxy for variations in the precipitation:evaporation ratio, which reflect first order hydrological variability. The results of this verification support our hypothesis that magnetic susceptibility records regional hydrological change as related to precipitation and lake level. Using both proxy data, we analyzed the past 3800 calendar years of hydrological variability. Our analyses indicate a long period of dry, less variable climate between 3800 and 2000 calendar years B.P. followed by a wet, more variable climate to the present. These results suggest that droughts of greater magnitude and duration than those observed in the modern record have occurred in the recent geological past. This conclusion presents insight to the potential impact of future droughts on the over-populated, water-poor region of Southern California.

## Introduction

There exist only a few natural, permanent lakes of Holocene age in coastal southwestern North America. These lakes are subject to frequent lake level change related to natural variations in precipitation (Lynch 1931; French and Busby 1974; Kirby et al. 2002a, b). The more numerous

Holocene playa lakes of the Mojave Desert region are characterized conversely by long, dry intervals punctuated by infrequent and short-lived wet periods (Ore and Warren 1971; Cole and Webb 1985; Enzel et al. 1989, 1992; Rosen 1991; Brown and Rosen 1995). Consequently, Holocene-aged playa lake sediment records from the Mojave region are incomplete. Acquisition of continuous Holocene-age sediment records from the natural permanent lakes of the coastal southwest will provide critical insight to the natural dynamics of long-term precipitation variability for coastal southwestern North America.

It is common knowledge that the growing population of coastal southwestern North America is threatened by a freshwater crisis (Changnon 2000; Metropolitan Water District 2003). At present, >50% of California's total population resides in the coastal southwest; yet, this region receives less than 2% of the state's total precipitation and less than 0.02% of the state's stream flow (USDA Forest Service: Pacific Southwest Region website [www.r5.fs.fed.us/water_resources/index.htm]). To account for the gross disparity between water supply and demand, a copious volume of freshwater is commuted to the coastal southwest via a series of aqueducts. Nevertheless, there remains a significant concern for future water shortages (i.e., droughts, Metropolitan Water District 2003). Further, with global climate models predicting both warming and more extreme variations in precipitation, the future water supply could be ever more precarious (IPCC 2001). It is therefore critical that scientists establish climatic 'baselines' for comparison of past climate to present and future climate. It is from these comparisons that policy can be instituted for preparing for potential 'mega-droughts' and their socio-economic impacts.

Here we present the initial sedimentological and geochemical records from a natural, permanent lake in coastal southwestern North America: Lake Elsinore. We assess the twentieth century relationship between lake level/precipitation variability and magnetic susceptibility from a lake basin sediment core. Based on the twentieth-century relationship between lake level and magnetic susceptibility, it is our contention that variations in magnetic susceptibility reflect lake level dynamics, which can be inferred over the late Holocene (past ~3800 calendar years B.P.). First order (decadal-to-millennial scale) changes in magnetic susceptibility reflect weathering/erosion cycles of the surrounding terrain in response to variations in precipitation (e.g., mid-Holocene Neoglacial transition). Application of these proxies to longer littoral cores provides a ~3800 year record of lake level dynamics (i.e., precipitation variability) that includes evidence for droughts on decadal and longer timescales. $\delta^{18}O_{(calcite)}$ data from lake sediment cores support our interpretation of magnetic susceptibility as a first-order proxy for lake level dynamics.

## Study site

Located 120 km SE of Los Angeles, California, Lake Elsinore is one of the few natural, permanent lakes in coastal southwestern North America (Figures 1 and 2). Lake Elsinore is a structural depression formed within a graben along the Elsinore fault (Mann 1956; Hull 1990). Geologically, Lake Elsinore is surrounded by a combination of igneous and metamorphic rocks, some of which outcrop in the lake's littoral zone along the northern edge (Engel 1959; Hull 1990). Lake Elsinore is constrained along its southern edge by the steep, deeply incised Elsinore Mountains that rise to more than 1000 m. The San Jacinto Mountains lie 70 km to the northeast of Lake Elsinore and rise to a maximum height of 3290 m; although, the land in between the San Jacinto Mountains and Lake Elsinore is characterized by a more gentle, hilly topography. Total sediment thickness underlying Lake Elsinore is estimated to be more than 600 m (Mann 1956; Pacific Groundwater Digest 1979; Damiata and Lee 1986; Hull 1990). Two exploratory wells have been drilled at the east end of the lake to 542 and 549 m, respectively, with sediment described as mostly fine-grained (Pacific Groundwater Digest 1979).

Lake Elsinore has a relatively small drainage basin (<1240 km$^2$) from which the San Jacinto River flows (semi-annually) into and terminates within the lake's basin (USGS 1998). An analysis of San Jacinto River discharge near San Jacinto, CA over the interval 1920 to 2001 A.D. illustrates the river's ephemeral behavior with discharge limited to an annual average of 0.56 m$^3$/s, 70% of which occurs during the months of February,

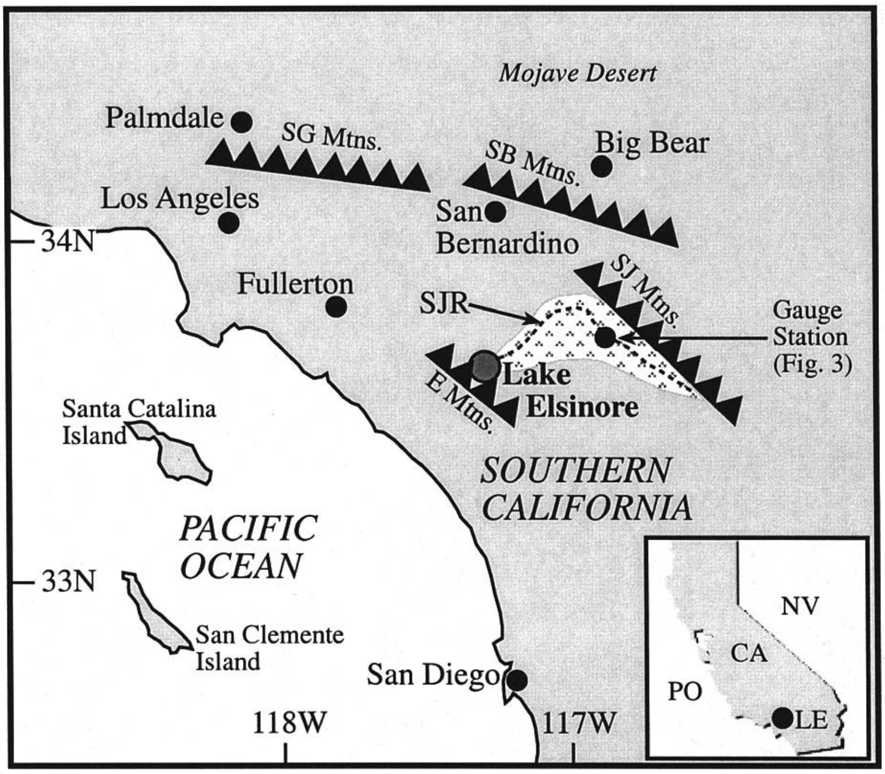


*Figure 1.* Site location map showing relevant regional information. Inset of California (CA), Nevada (NV), and the Pacific Ocean (PO) shown in lower right corner. SG – San Gabriel Mountains; SB – San Bernardino Mountains; SJ – San Jacinto Mountains; E – Elsinore Mountains; SJR – San Jacinto River. Area with pattern indicates the Lake Elsinore drainage basin during flow of the San Jacinto River. Location of San Jacinto River gauge station (Figure 3) is shown.

March, and April (Figures 1 and 3, http://waterdata.usgs.gov). The San Jacinto River originates from within the San Jacinto Mountains (Figure 1). As previously noted, these mountains achieve a maximum height of 3290 m, approximately 1000 m above the average annual snowline elevation of 2300 m in the adjacent San Bernardino Mountains (Figure 1, Minnich 1986). Consequently, it is possible that low $\delta^{18}O$ spring run-off impacts the lake's hydrology, and thus the $\delta^{18}O_{(calcite)}$ signal, via the San Jacinto River during above average snow accumulation winters or during long intervals of a cool, wet climate. Lake water $\delta^{18}O$ data, however, collected over the exceptionally wet hydrological year of 1993 do not indicate a significant 'alpine' affect (Williams and Rodoni 1997).

A small reservoir north of Lake Elsinore (Railroad Canyon Reservoir: RRCR) was built in 1927 as an attempt to moderate Lake Elsinore's lake level and supply water for surrounding agriculture (Mann 1947). The capacity of RRCR is small compared to the total capacity of Lake Elsinore (~12%) and thus it has proven rather ineffective in controlling lake level change (Mann 1947). As we show in the discussion section, the construction of RRCR has neither masked nor muted Lake Elsinore's first-order response to variations in precipitation. Construction of RRCR may also impact lake sedimentation dynamics. Here again we argue that construction of RRCR has not significantly affected the lake's first order response to climate-forced sedimentological change (see Discussion). In a continued effort to control lake level, the City of Lake Elsinore constructed a causeway in 1990 A.D. to reduce the lake's natural size, thus reducing evaporative losses. The causeway is shown on Figure 2 as the narrow white band and T-shaped protuberance that stretches across the lake near the southeast end. The impact of the causeway is not considered in

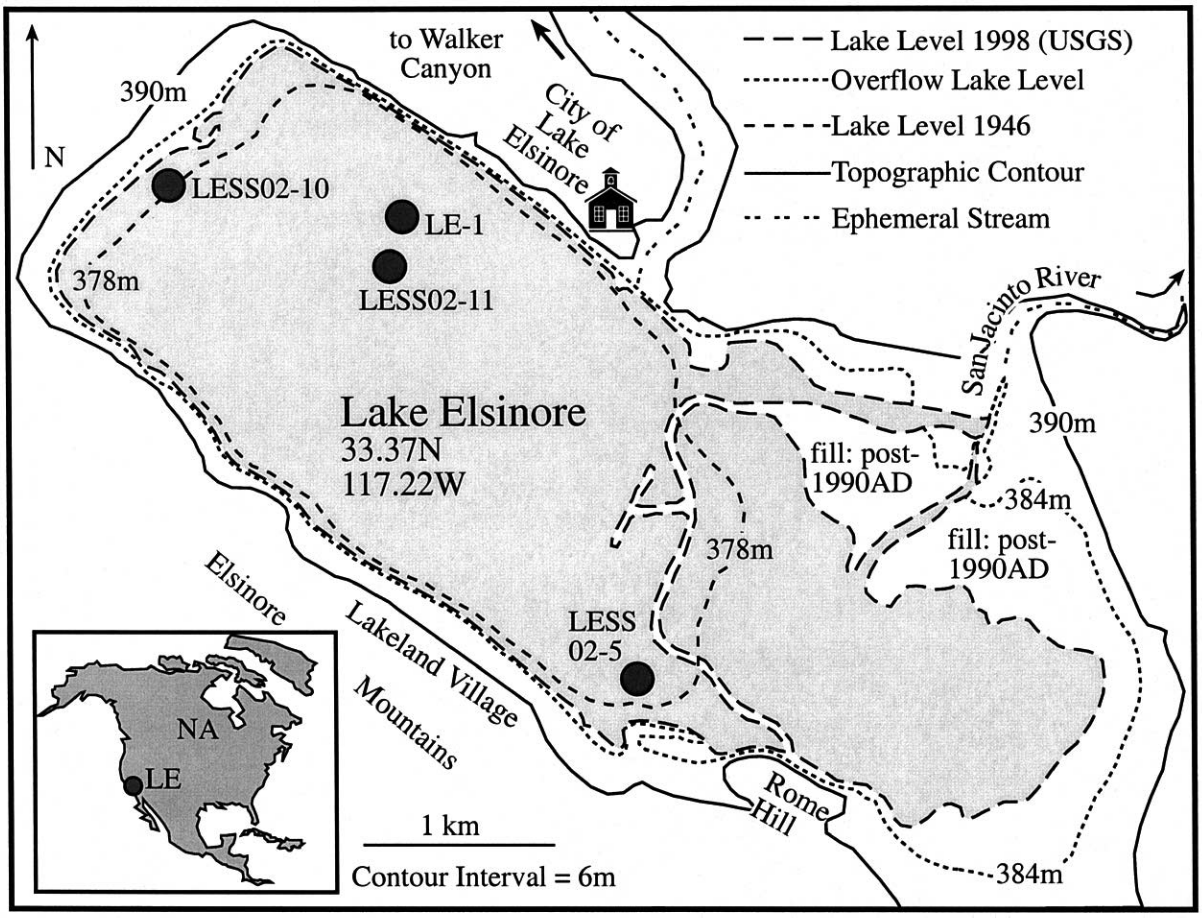


*Figure 2.* Lake Elsinore map with various relevant contours. Present day position of the lake shore is slightly below the 378 m contour. Inset at lower left of figure shows Lake Elsinore (LE) location relative to North America (NA).

this paper because it post-dates nearly the entire sediment record of interest.

Lake Elsinore has overflowed to the northwest through Walker Canyon very rarely, only three times in the twentieth century and 20 times since 1769 A.D. (Lynch 1931; USGS Lake level Data, 2002). Each overflow event was very short-lived (<several weeks) demonstrating that Lake Elsinore is essentially a closed-basin lake system (Lynch 1931; USGS Lake level Data, 2002). Conversely, Lake Elsinore has dried completely on only four occasions since 1769 A.D. (Lynch 1931; USGS Lake level Data, 1998). During periods of lake low stands, the lake is described by locals as 'nothing more than a marshy patch of tules' (perennial grasses) (Mann 1947). Interestingly, our initial sedimentological data from cores extracted from the deepest part of the profundal zone show no obvious evidence for periods of non-deposition (i.e., sediment hiatuses); however, our results indicate possible sediment hiatuses in the upper section of cores from the littoral environment. Archaeological evidence from continuous near shore sediment deposits indicates that Lake Elsinore contained water nearly continuously over the past 8400 years (dated via the $^{14}C$ method), permitting humans to thrive permanently within the area since at least the mid-Holocene (Grenda 1997). Any long-term lake drying or sustained flooding would have caused social upheaval and the subsequent abandonment of the near shore occupation site.

Limnologically, Lake Elsinore is a shallow, polymictic lake (13 m maximum depth based on historic

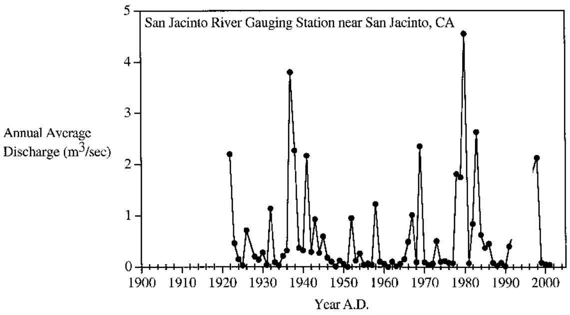


*Figure 3.* San Jacinto River gauging station near San Jacinto, CA (http://waterdata.usgs.gov). Note the large range of flow variability.

records) (Anderson 2001). A recent study by Anderson (2001) indicates that the hypolimnion is subject to short-lived periods of anoxia (i.e., days to weeks); although, the frequent mixing of oxygen rich epilimnion waters into the hypolimnion precludes permanent, sustained anoxia, at least during the period of observation. Annual water loss to evaporation from the lake's surface is >1.4 m; consequently, water residence time in Lake Elsinore is projected to be very short (<5 years?), and likely much shorter during drought periods (<1 year) (Mann 1947; USGS 1998; Anderson 2001). A strong grain size gradient exists from the littoral zone (i.e., coarse grained) to the profundal zone (e.g., fine grained). Mann (1947) attributes the grain size gradient to wave action winnowing and re-suspension of the finer-grained component of the littoral sediments into the profundal environment. This process of wave action winnowing and sediment re-suspension is a common and well-documented occurrence in most lake settings (Lehman 1975; Davis and Ford 1982; Downing and Rath 1988; Benson et al. 2002; Gilbert 2003; Miguel et al. 2003; Smoot 2003). Sediment trap studies indicate that $CaCO_3$ is produced within the water column, likely linked to photosynthetic uptake of $CO_2$ by phytoplankton (Anderson 2001). SEM analyses of lake sediment show distinct micron size $CaCO_3$ grains dispersed throughout the sediment (Anderson 2001).

Climatologically, coastal southwest North America is dominated by winter season (October to March) precipitation, which accounts for up to 80% of the annual water budget (Lynch 1931; USGS 1998; Redmond and Koch 1991; Friedman et al. 1992). At Lake Elsinore, the percent of winter precipitation for the months of December through February accounts for 60% of the annual total precipitation (NCDC Weather Station Locator Data [lwf.ncdc.noaa.gov/oa/climate/stationlocator.html]). The amount of winter precipitation is a function of the average position of the winter season polar front, which is forced by changes in the position of the eastern Pacific subtropical high, the polar front jet stream, and its modulation of storm tracks (Weaver 1962; Pyke 1972; Minnich 1984; Lau 1988; Schonher and Nicholson 1989; Enzel et al. 1989, 1992; Redmond and Koch 1991; Friedman et al. 1992; Ely 1997). In turn, large-scale atmospheric patterns that determine the average position of the polar front are modulated by Pacific Ocean sea-surface conditions (Namias 1951; Namias and Cayan 1981; Douglas et al. 1982; Lau 1988; Namias et al. 1988; Latif and Barnett 1994; Trenberth and Hurrell 1994; Cayan et al. 1998; Dettinger et al. 1998). As a result, Lake Elsinore lake level is controlled principally by the amount of winter precipitation received over its drainage basin, in turn reflecting large-scale ocean–atmosphere interactions over the Pacific Ocean (Kirby et al. 2002b).

## Age control

### *Age control for profundal cores*

A 270-year composite sediment core chronology for cores extracted from Lake Elsinore's deepest basin (LESS02-11 and LE-1) was constructed using several dating methods, including: non-native pollen types, changes in elemental lead concentrations, $^{137}Cs$ activity, and accelerator mass spectrometer (AMS) $^{14}C$ dating.

Two cores were used to construct a composite age model for the past 270 years: cores LE-1 and core LESS02-11 (Figure 2). Due to the difficultly in obtaining sediment cores longer than 180 cm, neither core by itself spans the past 270 years in its entirety. Core LE-1 only recovered the upper 113 cm of sediment. Whereas, core LESS02-11 recovered 180 cm of sediment, but over two separate drives. As a result, core LESS02-11 is missing sediment from 85 to 110 cm depth. To

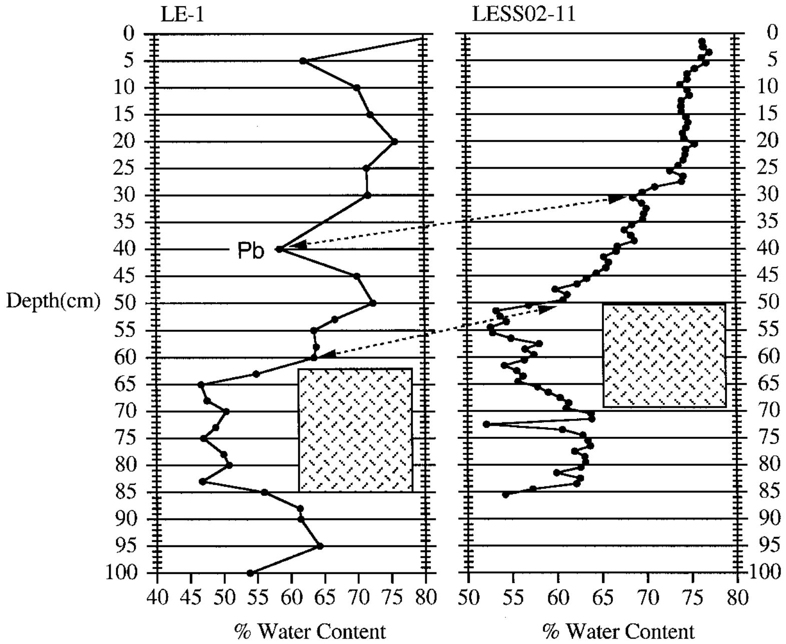


*Figure 4.* Percent water content graphs for cores LE-1 and LESS02-11. Location of elemental Pb peak is shown in core LE-1. Correlation of Pb depth to core LESS02-11 is shown by dashed line with arrows. Low stand sediment facies shown by box with pattern. Dashed line with arrows shows the cross-correlation of the low stand sediment facies uppermost boundary between cores LE-1 and LESS02-11.

*Table 1.* Radiocarbon dates.

| Identification | Core depth (cm) | Conventional radiocarbon age | Calibrated age[a] (cal. years B.P.) | 2-Sigma calibrated age (cal, years B.P.) | $^{13}C/^{12}C$ |
|---|---|---|---|---|---|
| Beta-169437 | LESS02-5 (21–22 cm) | 2010 ± 40 BP | 1960 cal years B.P. | 2050 to 1880 cal. years B.P. | −23.1 |
| Beta-169438 | LESS02-5 (78–79 cm) | 3220 ± 40 BP | 3450 cal years B.P. | 3470 to 3390 cal years B.P. | −23.3 |
| Beta-171203 | LESS02-11 (core top) | 103.5 ± pMC | 0 cal years B.P. | NA | −22.8 |
| Beta-171204 | LESS02-11 (145 cm) | 810 ± 40 BP | 715 cal years B.P. | 745 to 685 cal years B.P. | −23.7 |

[a]Stuiver et al. (1998).

assess better how the sediment from the two cores cross-correlated, we used percent water content data as a quick and easy method of cross-correlation (see Methods). Both cores contain similar lithologies characterized by brown, homogeneous clay sediments. The only noticeable difference within the sediments of both cores is the presence of a stiff, crumbly clay interval between 65 and 84 cm in core LE-1 and between 50 and 70 cm in core LESS02-11. The percent water content data clearly demarcate this stiff, crumbly clay layer (Figure 4). In both cores, the percent water content achieves and maintains low values across the sediment interval (Figure 4). We use the similarity in percent water content between cores LE-1 and LESS02-11, specifically the low water content interval, as a tie point for cross-correlating the core stratigraphies and age data for the development of a composite age model (Figure 4).

Age control for LESS02-11 is based on AMS $^{14}C$ dates on total organic carbon (Table 1), a $^{137}Cs$ peak (1963 A.D.), the first appearance of *Eucalyptus* (1910 ± 10 A.D.), and the first appearance of *Erodium cicutarium* (1800 ± 20 A.D.) (Figure 5, Mensing and Byrne 1998). Surface sediments from LESS02-11, dated by the AMS $^{14}C$ method, indicate post-1950 A.D. sediment (i.e., modern sediment) (Figure 5 and Table 1).

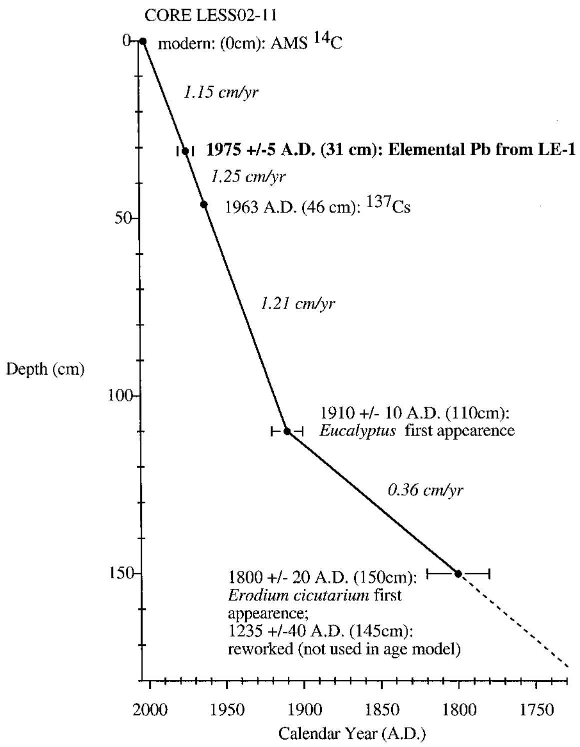


*Figure 5.* Age model for profundal core LESS02-11. Linear sedimentation rates are assumed between data points. Pb date from core LE-1 is shown by bold text (1975 ± 5 A.D.). LESS02-11 AMS $^{14}$C date calibrated to years A.D. (Stuiver et al. 1998).

The modern radiocarbon surface sediment age suggests that there is no 'old carbon' effect in the lake basin at present. The second AMS $^{14}$C date from LESS02-11 (715 cal year B.P.) is from just above (145 cm) the appearance of the *Erodium* pollen at 150 cm (Figure 5). The sediment from which the *Erodium* pollen and second AMS $^{14}$C date were determined is crumbly with a very low water content, similar to the crumbly clay sediment unit up-core in LESS02-11. This crumbly, low water content sediment facies is interpreted as an indicator of low lake levels. This sediment facies interpretation is supported by water content data from core LESS02-11 at the depth-age interval of the historic low lake level stand at Lake Elsinore between 1950 A.D. and 1965 A.D. (Figure 4 and 5). Because the $^{14}$C date comes from a sediment interval coeval to a lake level low stand, we consider this date to be too old, possibly due to reworked organic matter from littoral sediments during falling lake level (Smoot 2003). The pollen date, on the other hand, is likely derived via eolian processes and thus it is unlikely to be of reworked origin. The 'reworked' $^{14}$C date at 145 cm is not used in our age model.

LE-1 is dated from a peak in elemental Pb concentration associated with peak usage of leaded gasoline in the United States circa 1975 ± 5 A.D. (Figure 5, Callender and Rice 2000).

Using the water content data from cores LESS02-11 and LE-1 as the basis for cross-correlation, the elemental Pb data from core LE-1 was correlated to core LESS02-11 for the construction of a single composite core for use in developing a composite age model for the past 270 years (Figures 4 and 5).

### *Age control for littoral cores*

Age control for LESS02-5 is based on 2 AMS $^{14}$C dates on total organic carbon (Figure 6 and Table 1). The upper date (21.5 cm = 1960 calendar years B.P. [2010 ± 40 $^{14}$C years B.P.] see Table 1) was taken from the base of a rapid change in magnetic susceptibility from lower to higher values (Figure 7). The lower date (78.5 cm = 3450 calendar years B.P. [3220 ± 40 $^{14}$C years B.P.] see Table 1) was taken from a peak in a broad magnetic susceptibility high (Figure 7). The upper 7–8 cm of core LESS02-5 is dated by isotope stratigraphy (Figure 6). The values of $\delta^{13}C_{(calcite)}$ in core LESS02-5 increase dramatically between 7 and 8 cm (~1.5‰, Figure 7). This increase in $\delta^{13}C_{(calcite)}$ values is interpreted as an indicator of human development of the lake's drainage basin. As humans develop the drainage basin, the influx of nutrients into the lake increases. As a result, lake primary productivity increases extracting $^{12}$C from the lake dissolved inorganic carbon pool. Preferential removal of $^{12}$C by the primary producers causes the DIC pool in which the calcite precipitates to have higher $\delta^{13}$C values. Consequently, the $\delta^{13}C_{(calcite)}$ values are similarly higher. Alternatively, the $\delta^{13}C_{(DIC)}$ values of the lake water may have changed in response to regional crop development and tilling practices. Using this increase in $\delta^{13}C_{(calcite)}$ values between 7 and 8 cm as an indicator of drainage basin development, we assign a date of ~1880 A.D./~95 calendar years B.P. (late 19th century) to this depth interval (Figure 6). This age is approximately coeval to the initial development of the City of Lake Elsinore, both its infra-structure and its agricultural

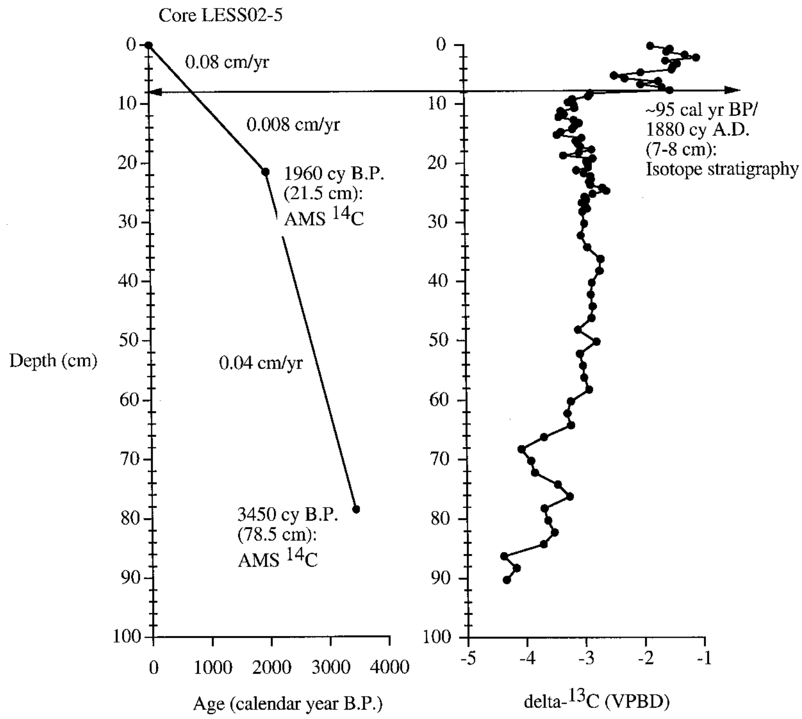


*Figure 6.* Age model for core LESS02-5. Linear sedimentation rates are assumed between data points. All AMS $^{14}C$ dates from core LESS02-5 are calibrated years B.P. (Stuiver et al. 1998).

development (McGlashan 1921). A 3-fold increase in the sedimentation rate of core LESS02-11 after 1910 A.D. support our argument that significant changes in the drainage basin have occurred, specifically changes in the flux of sediment into the lake basin (Figure 5). This increase in $\delta^{13}C_{(calcite)}$ values and sedimentation rates in response to anthropogenic effects on lake productivity and its drainage basin is recognized in a variety of lake systems worldwide (e.g., Schelske and Hodell 1991; Hodell et al. 1998; Teranes et al. 1999; Hilfinger et al. 2001; Meyers 2002). LESS02-10 is dated by cross-correlation to LESS02-5 using magnetic susceptibility stratigraphy (Figure 7).

## Methods

### *Core collection*

Two sediment cores (LESS02-11 and LE-1) were extracted from the profundal environment of Lake Elsinore using a square-rod piston corer (Figure 2). Core LE-1 (113 cm; $z$ = 5.6 m) was collected in Fall, 2001, core LESS02-11 (176 cm; $z$ = 5.5 m) in Summer 2002. Two additional cores (LESS02-5 [91 cm] and LESS02-10 [602 cm]) were extracted from the littoral environment (above water level) using a hollow-stemmed auger drill core in Spring 2002 (Figure 2). All cores were split, described, and archived for future sampling.

### *Percent water content*

Three to five $cm^3$ of wet sediment was placed into a pre-weighed beaker. The wet sediment weight is measured. The sediment is then dried at 60 °C for 24 h. The dry sediment was re-weighed to determine the percent weight water loss to evaporation (i.e., percent water content). Percent water content was determined for core LE-01 at various intervals ranging from 2.5 to 5.0 cm. Percent water content was determined for LESS02-11 at 1.0 cm intervals.

### *Magnetic susceptibility*

Samples were extracted from LESS02-5, LESS02-10, and LESS02-11 at 0.5 cm intervals. The samples

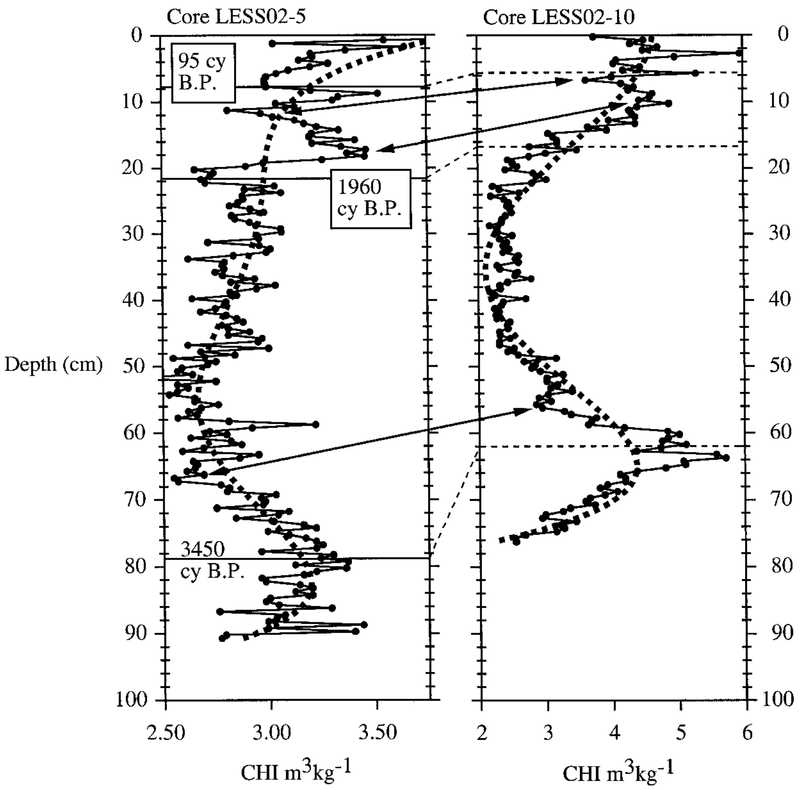


*Figure 7.* Environmental magnetic susceptibility data (CHI) for core LESS02-5 and -10. Solid lines represent absolute dates from core LESS02-5; dashed line represent cross-correlated dates from core LESS02-5 to core LESS02-10. Ages from core LESS02-5 are cross-correlated to core LESS02-10 using magnetic susceptibility stratigraphy. First-order trends in magnetic susceptibility data are shown by thick, dashed line. Solid lines with arrows show points of correlation between core magnetic susceptibility data.

were placed in pre-weighed 10-cc plastic cubes and magnetic susceptibility was measured twice on each sample with the *y*-axis rotated once per analysis. All samples were analyzed using a Bartington MS2 Magnetic Susceptibility instrument. All magnetic susceptibility measurements were determined immediately (i.e., same day) after the core was split and described to avoid possible magnetic mineral diagenesis with exposure to air. Following the measurement of magnetic susceptibility, the samples were re-weighed to obtain total sediment weight. The average magnetic susceptibility value for each sample was then divided by the sample weight to account for mass differences. Measurements were made to the 0.1 decimal place and reported as mass magnetic susceptibility (CHI = $\chi$) in SI units ($\times 10^{-7}$ m$^3$ kg$^{-1}$).

## $\delta^{18}O_{(calcite)}$ and $\delta^{13}C_{(calcite)}$

Samples were extracted from LESS02-5 for $\delta^{18}O_{(calcite)}$ and $\delta^{13}C_{(calcite)}$ at 0.5-cm intervals between 0 and 27.75 cm, and at 2.0 cm between 28.25 and 90.25 cm. Analysis of sediments from LESS02-5 using coulometry indicated $CaCO_3$ contents between 6 and 22% (data not shown). Samples were dried at room temperature, loosely disaggregated, and sieved at <63 $\mu$m. Only sediment <63 $\mu$m were used for stable isotope analysis to avoid the influence of larger biogenic carbonates such as ostracods, gastropods, and bivalve carapaces (Kirby et al. 2002c). By examining the <63-$\mu$m grain size fraction only, a process tested by Kirby et al. (2002c), we are confident that the carbonate analyzed consists predominantly of calcite precipitated in the epilimnion without the significant influence of biogenic carbonates.

Carbonate samples were roasted *in vacuo* at 200 °C to remove water and volatile organic contaminants that may confound stable isotope values of carbonate. Stable isotope values were obtained using a Finnigan Kiel-III carbonate preparation device directly coupled to the inlet of a Finnigan MAT 252 ratio mass spectrometer in the stable isotope laboratories at Syracuse University and the University of Saskatchewan. Twenty to 40 $\mu$g of carbonate were reacted at 70 °C with three drops of anhydrous phosphoric acid for 90 s. Isotope ratios were corrected for acid fractionation and $^{17}$O contribution and reported in per mil notation relative to the VPDB standard. Precision and calibration of data were monitored through daily analysis of NBS-18 and NBS-19 carbonate standards. $\delta^{18}$Ovalues of the samples are bracketed by those of the standards. Precision is better than ±0.1‰ for both carbon and oxygen isotope values.

### *Meteorological indices and lake level data*

Meteorological data were obtained from the National Climatic Data Center weather observation station records. Total winter precipitation amounts were calculated for the months of December, January, and February. Years with more than 1 month missing were not included. Winter precipitation data from Los Angeles and San Diego were averaged per year, and they were converted to *z*-scores to standardize the data for comparison. All precipitation data were smoothed by a five-point moving average to remove interannual noise. Lake level data for Lake Elsinore were obtained from the United States Geological Survey (1998). Baldwin Lake level data were obtained from French and Busby (1974).

## Results

### *Core descriptions*

Core LESS02-11 and LE-1 consist of homogenous brown clay. The only noticeable difference within the sediments of both cores is the presence of a relatively stiff, crumbly clay interval between 65 and 84 cm in core LE-1 and between 50 and 70 cm and 140 and 161 cm in core LESS02-11.

Core LESS02-5 consists of four sediment intervals: (1) 0–10 cm: black, silty sand with visible organic detritus; (2) 10–68 cm: grey, silty sand grading down core into a grey sandy silt with some microfossils visible (i.e., gastopod shells); (3) 68–87 cm: grey clay with some sandy pockets and microfossils visible (i.e., gastropod shells); and, (4) grey, sandy silt with microfossils visible (i.e., gastropods). Core LESS02-10 also consists of four sediment intervals: (1) 0–14.5 cm: grey sand; (2) 14.5–18 cm: grey silt grading down core into fine silt; (3) 18–23.5 cm: grey–brown sand with some coarse gravel boarded above and below by sharp contacts; and, (4) 23.5–78 cm: grey sandy silt grading down core into a brown, fine silt.

### *Magnetic susceptibility*

LESS02-11 is characterized by $\chi$ values between ~4 and ~13 m$^3$ kg$^{-1}$ (Figure 8). $\chi$ values are low at the core top and rise abruptly between 6 and 12 cm; the $\chi$ values decrease from 12 cm to a low point at 65 cm core depth (Figure 8). A similar pattern is observed for the lower section of LESS02-11 from high values at 110 cm to a low point at 158 cm (Figure 8).

LESS02-5 and -10 are characterized by similar magnetic susceptibility stratigraphy, a broad, first-order sinusoidal curve from high to low to high $\chi$ values from core top to core bottom (Figure 7). $\chi$ values for both LESS02-5 and -10 range between ~2 and ~6 m$^3$ kg$^{-1}$ (Figure 7). The differences in the absolute values of $\chi$ between the LESS02-5 and -10 likely reflect the subtle vagaries of the local depositional environments. This difference is most prominent during intervals of high $\chi$ values where LESS02-10 is characterized by values twice the $\chi$ values of LESS02-5.

### $\delta^{18}O_{(calcite)}$ and $\delta^{13}C_{(calcite)}$

$\delta^{18}O_{(calcite)}$ values from LESS02-5 range between −2.8 and 1.2‰ VPDB (Figure 9). $\delta^{18}O_{(calcite)}$ values are relatively low between the core top and 20 cm (Figure 9). Between 20 cm and 22 cm, $\delta^{18}O_{(calcite)}$ values increase nearly 2.0‰ (Figure 9). $\delta^{18}O_{(calcite)}$ values remain uniformly high between 22 cm and 34 cm (highest $\delta^{18}O_{(calcite)}$ value is 1.2‰ at 34 cm)

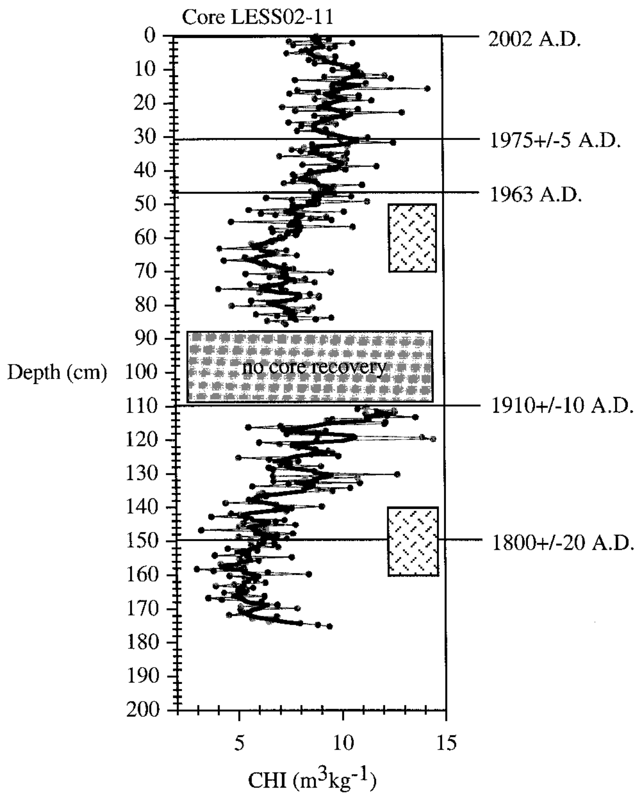


*Figure 8.* Environmental magnetic susceptibility data (CHI) for profundal core LESS02-11. Raw data shown by line with circles; five-point smoothed data shown by solid line. Box with pattern represent the low stand sediment facies.

after which they begin to decrease until reaching a minimum (−2.8‰) at 74 cm (Figure 9). From 74 cm to the core bottom, $\delta^{18}O_{(calcite)}$ values increase to 0.0‰ (Figure 9).

$\delta^{13}C_{(calcite)}$ values from core LESS02-5 range between −4.4 and −1.0‰ VPDB (Figure 9). The $\delta^{13}C_{(calcite)}$ data show a general increase in values from −4.4 to −2.2‰ over the length of the sediment record. The highest $\delta^{13}C_{(calcite)}$ value occurs at 2 cm core depth. A dramatic 1.5‰ increase in $\delta^{13}C_{(calcite)}$ values occurs between 7 and 8 cm depth (Figure 9).

### *Meteorological indices and lake level data*

Winter precipitation data for Los Angeles, San Diego, and Lake Elsinore are characterized by similar patterns of variability and a high degree of correlation ($r = 0.95$, Figure 10). These data indicate strong patterns of interannual-to-decadal scale precipitation variability as previously recognized by Cayan et al. (1998) and Dettinger et al. (1998). Likewise, the lake level data from Lake Elsinore and Baldwin Lake (located ~70 km northeast of Lake Elsinore at ~1700 m higher elevation) show similar patterns of variability (Figure 10). The dry interval between 1945 and 1965 A.D. shows prominently in both the precipitation and lake level records (Figure 10).

## Discussion

### *Historical calibration of lake level proxy data*

As previously illustrated, there is a strong positive relationship between regional precipitation and lake level at Lake Elsinore and Baldwin Lake (Figure 10). If Lake Elsinore's natural response to regional hydrodynamics is affected significantly by the construction of RRCR, it would be expected that the relationship between regional precipitation

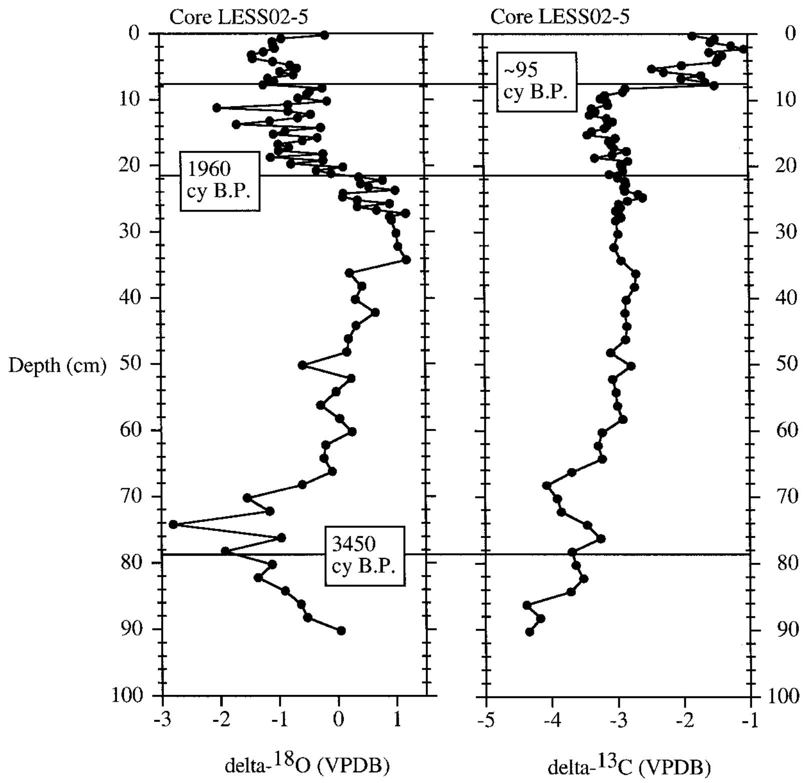


*Figure 9.* $\delta^{18}O_{(calcite)}$ and $\delta^{13}C_{(calcite)}$ data from core LESS02-5. Note the broad sinusoidal curve in the $\delta^{18}O_{(calcite)}$ data. Also, note the abrupt increase in $\delta^{13}C_{(calcite)}$ values are ~7–8 cm.

and Lake Elsinore lake level would become obfuscated since 1927 A.D.; clearly, this is not observed (Figure 10). Furthermore, Baldwin Lake, which is neither dammed nor developed by humans, shows the same response to regional precipitation as Lake Elsinore (Figure 10). As a result, it is concluded that both lake systems are responding to similar, regional precipitation variability unimpeded by human activity.

In developing a sediment proxy for lake level change, we conjectured that, in addition to increasing lake level, greater total precipitation should increase the amount of weathering and erosion in the lake's drainage basin. As weathering and erosion increase, the supply of detritus to the lake's profundal environment increases. As a result, the flux of magnetic minerals into the lake's basin also increases. The relationship between drainage basin weathering, sediment supply, and magnetic susceptibility has been shown by several researchers (Thompson et al. 1975; Bjorck et al. 1982; Brown et al. 2002; Yu and Kelts 2002). Using LESS02-11 from the lake's center and deepest basin, we measured environmental magnetic susceptibility ($\chi$) at 0.5-cm intervals over the past ~270 years (Figure 11). This age estimate is based on the composite age model from LE-1 and LESS02-11 (Figure 5). It is our working hypothesis that the values of environmental magnetic susceptibility from Lake Elsinore sediments are a reliable, first order proxy for lake level change. To test this hypothesis, we compared the twentieth-century plots of $\chi$ from LESS02-11 and Lake Elsinore lake level (Figure 11). Through this comparison, it is apparent that *first-order* lake level changes are recorded by magnetic susceptibility: higher lake levels correspond to higher $\chi$ values and vice versa (Figure 11). Because lake sedimentation is generally non-linear and site dependent, we do not assign a continuous age model to core LESS02-11. As a result, the relationship between lake level and $\chi$ is not a perfect match. Nonetheless, the first-order change from high to low to high lake levels over the twentieth century

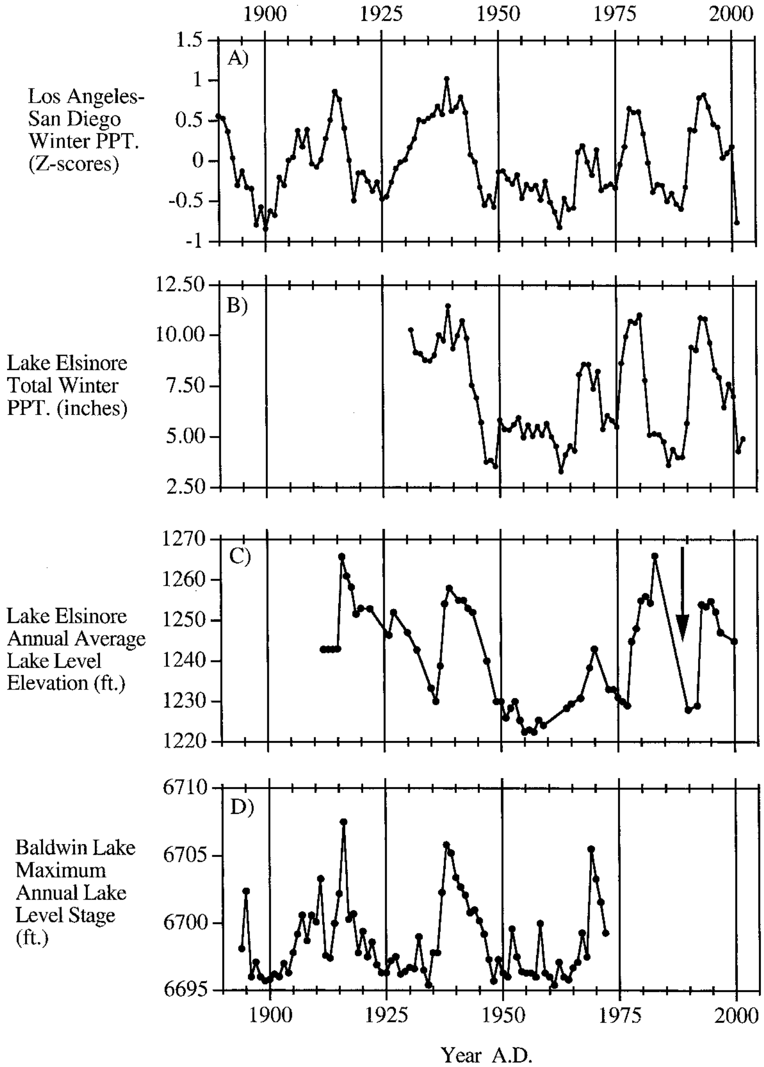


*Figure 10.* Graphs depicting winter precipitation and lake level data. (A) Los Angeles–San Diego winter precipitation (December–February) averaged and converted to *z*-scores. (B) Lake Elsinore total winter precipitation (December to February). All precipitation data are smoothed by a five-point moving average. Note the strong similarities between the regional precipitation data (A) and the local precipitation data (Lake Elsinore (B)). (C) Lake Elsinore annual average lake level elevation. Note the several years are missing data (e.g., 1960–63, 1984–89, etc.). (D) Lake Baldwin maximum annual lake level stage. Note the strong correlation between lake level change at both lake sites and winter precipitation.

is recorded by the $\chi$ data (Figure 11). Furthermore, only the most sustained intervals of low lake level (i.e., 1950–1965 A.D.) are recorded by continuously low $\chi$ data (Figure 11). Notably, the drought interval of the 1980s A.D. is not recorded by $\chi$ data (Figure 11). This observation leads us to conclude that there is a sediment threshold response to climate in Lake Elsinore. In other words, the $\chi$-climate connection is limited to a first-order response of time-dependent duration. Although this calibration of proxy data is based on the twentieth century only, we assume uniformity of this first order relationship over time, specifically over decadal-to-millennial time scales.

But, can we assume sediment-climate process uniformity if there exists a dam? Construction of RRCR, like most reservoirs, acts a large sediment trap for sediment derived from its dammed river, in

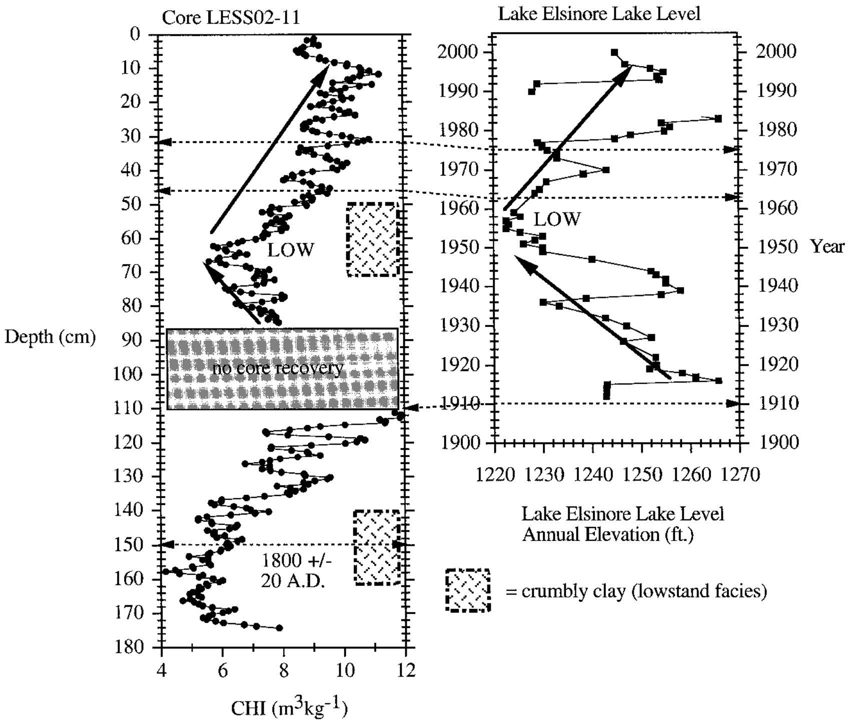


*Figure 11.* Left: Environmental magnetic susceptibility for core LESS02-11. Data are smoothed by a 5-point moving average. Absolute dates are noted with dashed lines with arrows. Solid lines with arrows represent similar trends or features between the CHI data and the lake level data. Right: Lake Elsinore annual average lake level elevation. Years without squares represent years without lake level data. Note the similar first order twentieth century patterns of change between CHI and lake level (i.e., low CHI = low lake level and vice versa).

this case the San Jacinto River. If the San Jacinto River was the only source of allochthonous sediment to Lake Elsinore, it would be difficult to assume first-order uniformity between climate (i.e., precipitation) and sedimentation over time. One look at the steep, deeply incised Elsinore Mountains clearly illustrates that the San Jacinto River is not the only sediment source for Lake Elsinore. In fact, direct run-off from the Elsinore Mountains may be a more important source of allochthonous sediment when averaged over an interval of a year and longer, particularly when you consider the ephemeral discharge of the San Jacinto River (Figure 3). Only twice between 1940 and 1965 A.D. did the San Jacinto River exceed an annual discharge of 1.0 $m^3/s$ (Figure 3). In other words, the San Jacinto River is more likely a periodic source of pulsed sediment whereas the Elsinore Mountains are a more consistent, or time-averaged, sediment source. As a result, it is concluded that the construction of RRCR is unlikely to have eliminated the sediment's response to climate in Lake Elsinore.

### *Application of historically calibrated lake level proxy to the prehistoric sediment record*

To apply our hypothesis linking lake level and $\chi$ values to the prehistoric sediment record, we examined the $\chi$ time series from LESS02-5 and -10 (Figure 7). LESS02-5 and -10 are located at opposite ends of the lake basin, yet there is a high degree of first-order coherency between their respective magnetic susceptibility records suggesting that both locations record similar sediment signals (Figure 7). Magnetic susceptibility data from a lake transect of 11 gravity cores confirm our observation of spatial coherency of the sediment magnetic stratigraphy (data not shown). The magnetic susceptibility records are characterized by first-order curves showing high-to-low-to-high $\chi$ values (Figure 7). Using the historically calibrated lake

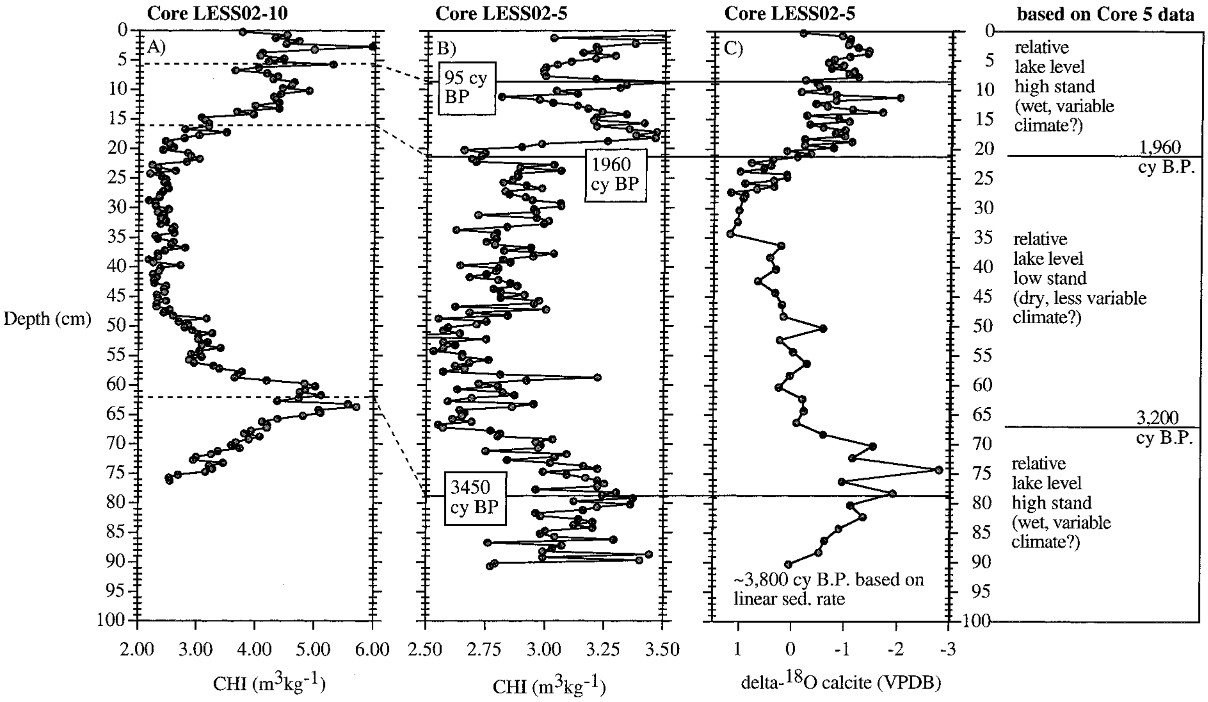


*Figure 12.* Summary figure. (A) Environmental magnetic susceptibility for core LESS02-10. (B) Environmental magnetic susceptibility for core LESS02-5. (C) $\delta^{18}O_{(calcite)}$ data for core LESS02-5. Solid lines represent absolute dates on core LESS02-5. Dates are correlated to core LESS02-10 via the magnetic susceptibility stratigraphy (dashed line). Note that both CHI and $\delta^{18}O_{(calcite)}$ show similar first order patterns of change in terms of inferred lake hydrodynamics.

level proxy, $\chi$, the sediments indicate three intervals of changing lake levels: (1) relative high lake levels from ~3800 (core bottom) to ~3200 calendar years B.P. (based on a linear age models between AMS $^{14}C$ dates [calibrated to calendar years B.P.] to sediment core bottom); (2) an interval of low lake levels between ~3200 and 1960 calendar years B.P.; and (3) relative high lake levels between 1960 calendar years B.P. and today (Figure 12). At face value, the sediment accumulation rates in the upper section (1960 to 95 cy B.P.) of LESS02-5, not including the twentieth century, are lower (0.008 cm/year assuming no sediment hiatuses) than during the preceeding interval (0.04 cm/year) of relative low lake levels (Figure 6). This observation is counterintuitive to our hypothesis that high lake levels correspond to more precipitation and thus an increase in the flux of eroded sediment into the lake basin. A likely explanation for this inconsistency is that there may be a gap, possibly several (?), in the upper section of littoral core LESS02-5 (and -10).

*Supporting proxy data for lake level change*

$\delta^{18}O$ values of inorganic lacustrine calcite are a well-documented climate proxy for lake studies (Stuiver 1970; McKenzie and Hollander 1993; Drummond et al. 1995; Anderson et al. 1997; Yu et al. 1997; Teranes and McKenzie 2001; Hammarlund et al. 2002; Kirby et al. 2002c, d). In arid lake environments, such as Lake Elsinore, there are multiple studies purporting traditional interpretations of $\delta^{18}O_{(calcite)}$ values in terms of climate-forced relative lake level change (Benson et al. 1991, 1996, 1998, 2002; Li and Ku 1997; Li et al. 2000; Seltzer et al. 2000). Although the absolute value of lake level cannot be discerned from $\delta^{18}O_{(calcite)}$ values, it is possible to infer the general climate regime in terms of precipitation (inflow) *versus* evaporation (outflow) (P:E ratio) (Benson et al. 1991; Johnson et al. 1991; Lister et al. 1991; Li and Ku 1997; Li et al. 2000; Seltzer et al. 2000; Benson and Paillet 2002; Benson et al. 2002). Relatively dry climates (lower relative lake levels, P:E $< 1.0$) favor the evaporation of water from the lake basin preferentially removing $^{16}O$ from the lake water. As a result, the $\delta^{18}O_{(calcite)}$ values increase during dry climates. Relatively wet climates (higher relative lake levels) are characterized by less evaporation (P:E $> 1$) of water from the lake basin compared to the influx of precipitation and its run-off. As a result, the $\delta^{18}O_{(calcite)}$ values are lower during wet climates. During wetter, cooler climates, the increased contribution of $\delta^{18}O$ winter precipitation/run-off (lower $\delta^{18}O_{(precipitation)}$ values) into the lake basin will help to additionally lower the $\delta^{18}O_{(calcite)}$ values. The seasonal variations, however, in $\delta^{18}O_{(precipitation)}$ should be minimal when compared to the annual effect of evaporation on $\delta^{18}O_{(lake\ water)}$ values, particularly for Lake Elsinore where the seasonal distribution of precipitation is limited to one season only (Friedman et al. 1992; Williams and Rodoni 1997). $\delta^{18}O_{(calcite)}$ also tend to conflate temperature variations. Given that our $\delta^{18}O_{(calcite)}$ values display a 4‰ range (−3.0 to >1.0‰), it is unlikely that changes in temperature (>12 °C) could be solely or even mainly responsible for the observed variations in $\delta^{18}O_{(calcite)}$. We conclude, therefore, that the $\delta^{18}O_{(calcite)}$ data predominantly reflect changes in the P:E ratio with changes in lake water temperature and variations in $\delta^{18}O_{(precipitation)}$ as secondary influences. Our interpretation of $\delta^{18}O_{(calcite)}$ values in terms of lake-climate information is congruent with most arid environment lake studies (e.g., Benson et al. 1991; Johnson et al. 1991; Lister et al. 1991; Li and Ku 1997; Seltzer et al. 2000; Benson et al. 2002).

Using this basic interpretation of $\delta^{18}O_{(calcite)}$ in terms of P:E changes, we compare the $\delta^{18}O_{(calcite)}$ record from core LESS02-5 to the $\chi$ record from core LESS02-5 (Figure 10). The interpretation of both records is similar in terms of climate-forced relative lake level changes (Figure 10). From ~3800 (the limit of the sediment record) to ~3200, lake level was relatively high, or the P:E ratio was greater than 1.0 (i.e., precipitation/run-off exceeded evaporation) (Figure 10). From ~3300 to 1960 calendar years B.P. lake level was relatively low (Figure 10). The presence of sediment during this low interval precludes complete desiccation of the lake basin and the erosion and/or non-deposition of sediment for any significant length of time. Thus, we interpret this climate as dry, but less variable wherein the climate became more arid but eventually established a dynamic equilibrium between precipitation/run-off and evaporation, thus preventing the complete drying of the lake basin.

The upper sediment section is more difficult to interpret. Straightforward interpretation of the proxy data suggests that the upper section represents an interval of relatively high lake levels; however, the low sedimentation rates suggest that a large sediment hiatus may exist. It is unlikely that sedimentation rates are lower during high stands than during low stands; although, recent lake sediment research by Smoot (2003) may suggest otherwise. Without additional dates, however, it is not presently possible to determine the location and duration of the sediment hiatus. The presence of significant hiatuses during our interpreted wet interval (i.e., higher lake levels) suggests that the climate was wet, but more variable perhaps with high amplitude climate variations. This period of high amplitude climate variability was likely characterized by fluctuations between extreme wet and extreme dry climates. Dry climates likely produced brief but significant droughts. These droughts would have desiccated the lake creating the suspected sediment hiatuses via erosion of the littoral environment. Based on our research, southern California presently resides in a wet, but highly dynamic climate state.

## Conclusion

An analysis of sediment from Lake Elsinore in coastal southwestern North America indicates dynamic lake level variability. The premise for the inferred lake level variability is a historically calibrated relationship between lake level, regional precipitation, and magnetic susceptibility. This relationship indicates that first order variations in magnetic susceptibility are positively correlated to lake level fluctuations via regional precipitation. A weathering/erosion hypothesis is proposed to explain this observation. Using the historically calibrated lake level proxy (i.e., magnetic susceptibility), the past 3800 calendar years are interpreted in terms of general climate regimes and associated relative lake levels. As an independent proxy for lake hydrodynamics, we used $\delta^{18}O_{(calcite)}$ to test our interpretation of the magnetic susceptibility data. The $\delta^{18}O_{(calcite)}$ data verify our hypothesis that magnetic susceptibility records lake level change. Using these two proxies, three distinct intervals of relative lake level and general climate state are inferred: (1) relative high lake levels (wet, variable climate) from 3800 cal year B.P. to 3200 cal year B.P.; (2) relative low lake levels (dry, less variable climate) from 3200 cal year B.P. to 1960 cal year B.P.; and (3) relative high lake levels (wet, variable climate) from 1960 cal year B.P. through the present. These results suggest that mega-scale droughts (decadal-to-centennial scale) of greater magnitude and duration than observed in the historical record have occurred in coastal southwestern North America in the recent geological past. Population growth and its demands on freshwater usage will exacerbate the impact of future droughts in Southern California.

## Acknowledgements

This research was funded by the Santa Ana Regional Water Quality Control Board, Cal-State Fullerton, the University of Southern California, the University of Saskatchewan, and the University of California at Berkeley. Special thanks to Mr. Patrick Kilroy (Manager of Lake Elsinore), Dr. Xinyu (Cindy) Li (State Water Board's Contract Manager), the City of Lake Elsinore, and Mr. Dennis Waite (Boat Technician) for site access, boat use, and assistance. Thoughtful reviews by Dr. Michael Talbot and Dr. Michael Rosen helped to clarify several important issues; their time is greatly appreciated.

## References

Anderson M.A. 2001. Internal Loading and Nutrient Cycling in Lake Elsinore: Final Report to Santa Ana Regional Water Quality Control Board, 52 pp.

Anderson W.T., Mullins H.T. and Ito E. 1997. Stable isotope record from Seneca Lake, New York: evidence for a cold paleoclimate following the Younger Dryas. Geology 25: 135–138.

Benson L.V., Meyers P.A. and Spencer R.J. 1991. Change in the size of Walker Lake during the past 5000 years. Palaeogeogr. Palaeoclimatol. Palaeoecol. 81: 189–214.

Benson L.V., Burdett J.W., Kashgarian M., Lund S.P., Phillips F.M. and Rye R.O. 1996. Climatic and hydrological oscillations in the Owens Lakes Basin and Adjacent Sierra Nevada, California. Science 274: 746–749.

Benson L.V., May H.M., Antweiler R.C., Brinton T.I., Kashgarian M., Smoot J.P. and Lund S.P. 1998. Continuous lake-sediment records of glaciation in the Sierra Nevada between 52,600 and 12,500 14C year B.P.. Quat. Res. 50: 113–127.

Benson L. and Paillet F. 2002. HIBAL: a hydrologic-isotopic-balance model for application to paleolake systems. Quat. Sci. Rev. 21: 1521–1539.

Benson L., Kashgarian M., Lund S., Paillet F., Smoot J., Kester C., Meko D., Lindstrom S., Mensing S. and Rye R. 2002. Multidecadal and multicentennial droughts affecting Northern California and Nevada: implication for the future of the West. Quat. Res. 21: 659–682.

Bjorck S., Dedaring J.A. and Jonsson A. 1982. Magnetic susceptibility of Late Weischselian deposits in southeastern Swedan. Boreas 11: 99–111.

Brown W.J. and Rosen M.R. 1995. Was there a Pliocene-Pleistocene fluvial-lacustrine connection between Death Valley and the Colorado River. Quat. Res. 43: 286–296.

Brown S., Bierman P., Lini A., Davis P.T. and Southon J. 2002. Reconstructing lake and drainage basin history using terrestrial sediment layers: analysis of cores from a post-glacial lake in New England, USA. J. Paleolim. 28: 219–236.

Callender E. and Rice K.C. 2000. The urban environmental gradient: Anthropogenic influences on the spatial and temporal distributions of lead and zinc in sediments. Environ. Sci. Technol. 34: 232–238.

Cayan D.R., Dettinger M.D., Diaz H.F. and Graham N.E. 1998. Decadal variability of precipitation over Western North America. J. Climate 11: 3148–3166.

Changnon S.A. 2000. Reactions and responses to recent urban droughts. Phys. Geogr. 21: 1–20.

Cole K.L. and Webb R.H. 1985. Late Holocene changes in Greenwater Valley, Mojave Desert, California. Quat. Res. 23: 227–235.

Damiata B.N. and Lee T.-C. 1986. Geothermal exploration in the vicinity of Lake Elsinore, Southern California. Geothermal Resource Council Trans. 10: 119–123.

Davis M.B. and Ford M.S.J. 1982. Sediment focusing in Mirror Lake, New Hampshire. Limnol. Oceanogr. 27: 137–150.

Dettinger M.D., Cayan D.R., Diaz H.F. and Meko D.M. 1998. North south precipitation patterns in western North America on interannaual-to-decadal timescales. J. Climate 11: 3095–3111.

Douglas A.V., Cayan D.R. and Namias J. 1982. Large-scale changes in North Pacific and North American weather pattern in recent decades. Monthly Weather Rev. 110: 1851–1862.

Downing J.A. and Rath L.C. 1988. Spatial patchiness in the lacustrine sedimentary environment. Limnol. Oceanogr. 33: 447–458.

Drummond C.N., Patterson W.P. and Walker J.C.G. 1995. Climatic forcing of carbon–oxygen isotopic covariance in temperate region marl lakes. Geology 23: 1031–1034.

Ely L.L. 1997. Response of extreme floods in the southwestern United States to climatic variations in the late Holocene. Geomorphology 19: 175–201.

Engel R. 1959. Geology of the Lake Elsinore quadrangle, California. Calif. Div. Mines Bull. 146: 52

Enzel Y., Cayan D.R., Anderson R.Y. and Wells S.G. 1989. Atmospheric circulation during Holocene lake stands in the Mojave Desert: evidence of regional climate change. Nature 341: 44–47.

Enzel Y., Brown W.J., Anderson R.Y., McFadden L.D. and Wells S.G. 1992. Short-duration Holocene lakes in the Mojave River drainage basin, Southern California. Quat. Res. 38: 60–73.

French J.J. and Busby M.W. 1974. Flood-hazard study – 100-year flood stage for Baldwin Lake San Bernardino County California. US Geological Survey Water-Resources Investigations 26–74, pp. 1–18.

Friedman I., Smith G.I., Gleason J.D., Warden A. and Harris J.M. 1992. Stable Isotope Composition of Waters in Southeastern California 1. Modern Precipitation. J. Geophys. Res. 97: 5795–5812.

Gilbert R. 2003. Spatially irregular sedimentation in a small, morphologically complex lake: implications for paleoenvironmental studies. J. Paleolim. 29: 209–220.

Grenda D. 1997. Continuity and Change: 8,500 Years of Lacustrine Adaptation on the Shores of Lake Elsinore. Statistical Re. Tech. Ser. 59: 324.

Hammarlund D., Barnekow L., Birks H.J.B., Buchardt B. and Edwards T.W.D. 2002. Holocene changes in atmospheric circulation recorded in the oxygen-isotope stratigraphy of lacustrine carbonates from northern Swedan. The Holocene 12: 339–351.

Hilfinger M.F. IV, Mullins H.T., Burnett A. and Kirby M.E. 2001. A 2500 year sediment record from Fayetteville Green Lake, New York: evidence for anthropogenic impacts and historic isotope shift. J. Paleolim. 26: 293–305.

Hodell D.A., Schelske C.L., Fahnenstiel G.L. and Robbins L.L. 1998. Biologically induced calcite precipitation and its isotopic composition in Lake Ontario. Limnol. Oceanogr. 43: 187–199.

Hull A.G. 1990. Seismotectonics of the Elsinore-Temecula trough, Elsinore fault zone, southern California. Ph.D. Thesis, University of California, Santa Barbara, pp. 233.

IPCC Working Group, 2001. Summary for policy makers. http://www.ipcc.ch.

Johnson T.C., Halfman J.D. and Showers W.J. 1991. Paleoclimate of the past 4000 years at Lake Turkana, Kenya, based on the isotopic composition of authigenic calcite. Palaeoclimatol. Palaeoceanogr. Palaeoecol. 85: 189–198.

Kirby M.E., Lund S.P., Poulsen C.P., Patterson W.P. and Burnett A.W. 2002. On the Use of Historic Atmosphere-Lake level Relationships for Reconstructing Stable Oxygen Isotope-Based Paleohydrology in Southern CA. American Geophysical Union Fall Meeting.

Kirby M.E., Lund S.P., Poulsen C.P., Patterson W.P. and Hartman C. 2002. Evidence for High Amplitude, High Frequency Holocene Paleohydrological Change From a Southern Californian Lacustrine Site (Lake Elsinore, Riverside County). Abstracts with Programs – Geological Society of America Annual Meeting.

Kirby M.E., Mullins H.T., Patterson W.P. and Burnett A.W. 2002c. Late glacial-holocene atmospheric circulation and precipitation in the northeast United States inferred from modern calibrated stable oxygen and carbon isotopes. Geol. Soc. Am. Bull. 114: 1326–1340.

Kirby M.E., Patterson W.P., Mullins H.T. and Burnett A.W. 2002d. Post-Younger Dryas Climate Interval Linked to Circumpolar Vortex Variability: Evidence from Fayetteville Green Lake, New York. Climate Dynamics 19: 321–330.

Latif M. and Barnett T.P. 1994. Causes of decadal climate variability over the North Pacific and North America. Science 266: 634–637.

Lau N.-C. 1988. Variability of the observed midlatitude storm tracks in relation to low-frequency changes in the circulation pattern. J. Atmos. Sci. 45: 2718–2743.

Lehman J.T. 1975. Reconstructing the rate of accumulation of lake sediment: the effect of sediment focusing. Quat. Res. 5: 541–550.

Li H.-C. and Ku T.-L. 1997. d13C-d18O covariance as a paleohydrological indicator for closed-basin lakes. Palaeogeogr. Palaeoclimatol. Palaeoecol. 133: 69–80.

Li H.-C., Bischoff J.L., Ku T.-L., Lund S.P. and Stott L.D. 2000. Climate variability in east-central California during the past 1000 years reflected by high-resolution geochemical and isotopic records from Owens Lake sediments. Quat. Res. 54: 189–197.

Lister G.S., Kelts K., Chen K.-Z., Yu J.-Q. and Niessen F. 1991. Lake Qinghai, China: closed-basin lake levels and the oxygen isotope record for ostracoda since the latest Pleistocene. Palaeoclimatol. Palaeoceanogr. Palaeoecol. 84: 141–162.

Lynch H.B. 1931. Rainfall and stream run-off in Southern California since 1769. The Metropolitan Water District of Southern California, Los Angeles, California, pp. 1–31.

Mann J.F. 1947. The Sediments of Lake Elsinore. MSc Thesis, University of Southern California, pp. 33.

Mann J.F. Jr. 1956. The origin of Elsinore Lake Basin. Bull. Southern California Acad. Sci. 55: 72–78.

McGlashan H.D. 1921. Surface water supply of the Pacific slope of Southern California. United States Geological Survey Water-Supply Paper 47: 321–324.

McKenzie J.A. and Hollander D.J. 1993. Oxygen-isotope: record in recent carbonate sediments from Lake Greifen, Switzerland (1750–1986): application of continental isotopic indicator for evaluation of changes in climate and atmospheric circulation patterns. Climate Change Continental Isotopic Records Geophys. Monogr. 78: 101–111.

Mensing S. and Byrne R. 1998. Pre-mission invasion of Erodium cicutarium in California. J. Biogeogr. 25: 757–762.

Metropolitan Water District, 2003. Report on the Metropolitan's Water Supply, pp. 29.

Meyers P.A. 2002. Evidence of mid-Holocene climate instability from variation in carbon burial in Seneca Lake, New York. J. Paleolim. 28: 237–244.

Miguel S., Bolivar J.P. and Garcia-Tenorio R. 2003. Mixing, sediment accumulation and focusing using $^{210}Pb$ and $^{137}Cs$. J. Paleolim. 29: 1–11.

Minnich R.A. 1984. Snow drifting and timberline dynamics on Mount San Gorgonio, California, USA. Arctic Alpine Res. 16: 395–412.

Minnich R.A. 1986. Snow levels and amounts in the mountains of Southern California. J. Hydrol. 89: 37–58.

Namias J. 1951. The Great Pacific Cyclone of Winter 1949–50: a case study in the evolution of climatic anomalies. J. Meteorol. 8: 251–261.

Namias J. and Cayan D.R. 1981. Large-scale air-sea interactions and short-period climatic fluctuations. Science 214: 869–876.

Namias J., Yuan X. and Cayan D.R. 1988. Persistence of North Pacific sea surface temperature and atmospheric flow patterns. J. Climate 1: 682–703.

Ore H.T. and Warren C.N. 1971. Late-Pleistocene-early Holocene geomorphic history of Lake Mojave, California. Geol. Soc. Am. Bull. 82: 2553–2562.

Pacific Groundwater Digest (Anonymous), 1979. Aquifer Exploration at Lake Elsinore. Pacific Groundwater Digest August: 10–12.

Pyke C.B. 1972. Some Meteorological Aspects of the Seasonal Distribution of Precipitation in the Western United States and Baja California. University of California Water Resources Center 139, 215 pp.

Redmond K.T. and Koch R.W. 1991. Surface climate and stream flow variability in the Western United States and their relationship to large-scale circulation indices. Water Resources Res. 27: 2381–2399.

Rosen M.R. 1991. Sedimentologic and geochemical constraints on the hydrologic evolution of Bristol Dry Lake Basin, California, USA. Palaeogeogr. Palaeoclimatol. Palaeoecol. 84: 229–257.

Schelske C. and Hodell D. 1991. Recent changes in productivity and climate of Lake Ontario detected by isotope analysis of sediments. Limnol. Oceanogr. 36: 961–975.

Schonher T. and Nicholson S.E. 1989. The Relationship Between California Rainfall and ENSO Events. J. Climate 2: 1258–1269.

Seltzer G., Rodbell D. and Burns S. 2000. Isotopic evidence for late Quaternary climatic change in tropical South America. Geology 28: 35–38.

Smoot J.P. 2003. Impact of sedimentation styles on paleoclimate proxies in late Pleistocene through Holocene lakes in the western US. Third International Limnogeology Congress Abstract Volume: p. 273.

Stuiver M. 1970. Oxygen and carbon isotope ratios of freshwater carbonates as climatic indicators. J. Geophys. Res. 75: 5247–5257.

Stuiver M., Reimer P.J., Bard E., Beck J.W., Burr G.S., Hughen K.A., Kromer B., McCormac F.G., v.d. Plicht J. and Spurk M. 1998. INTCAL98 radiocarbon age calibration, 24,000-0 cal B.P.. Radiocarbon 40: 1041–1083.

Teranes J.L., McKenzie J.A., Lotter A.F. and Sturm M. 1999. Stable isotope response to lake eutrophication: calibration of a high-resolution lacustrine sequence from Baldeggersee, Switzerland. Limnol. Oceanogr. 44: 320–333.

Teranes J.L. and McKenzie J.A. 2001. Lacustrine oxygen isotope record of twentieth-century climate change in central Europe: evaluation of climate controls on oxygen isotopes in precipitation. J. Paleolim. 26: 131–146.

Thompson R., Battarbee R.W., O'Sullivan P.E. and Oldfield F. 1975. Magnetic susceptibility of lake sediments. Limnol. Oceanogr. 20: 687–698.

Trenberth K.B. and Hurrell J.W. 1994. Decadal atmosphere-ocean variations in the Pacific. Climate Dynamics 9: 303–319.

USGS Special Report, 1998. National Water-Quality Assessment Program – Santa Ana Basin: http://water.wr.usgs.gov/sana_nawqa/index.html.

Weaver R.L. 1962. Meteorology of Hydrologically Critical Storms in California. Hydrometeorological Report No. 37, US Weather Bureau, Washington, DC, pp. 110.

Williams A.E. and Rodoni D.P. 1997. Regional isotope effects and application to hydrologic investigations in southwestern California. Water Resources Res. 33: 1721–1729.

Yu J.Q. and Kelts K.R. 2002. Abrupt changes in climatic conditions across the late-glacial/Holocene transition on the N.E. Tibet-Qinghai Plateau: evidence from Lake Qinghai, China. J. Paleolim. 28: 195–206.

Yu Z., McAndrews J.H. and Eicher U. 1997. Middle-Holocene dry climate caused by changes in atmospheric circulation patterns: Evidence from lake levels and stable isotopes. Geology 25: 251–254.