Hydrologic variability and the onset of modern El Niño–Southern Oscillation: a 19 250-year record from Lake Elsinore, southern California

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ABSTRACT: There are very few terrestrial palaeoclimate archives spanning the Last Glacial Maximum through the Holocene from coastal southern California. Yet, knowledge of past climate dynamics is critical for assessing present and future constraints on the region’s dwindling freshwater resources. We present initial results from two drill cores extracted from the present-day edge of Lake Elsinore, southern California’s largest natural lake. Using a multiproxy approach including lithologic description, mass magnetic susceptibility, LOI 550°C (total organic matter), and LOI 950°C (total carbonate), we infer first-order, long-term climate change over the past 19 250 calendar years. Furthermore, we suggest possible first-order forcing mechanisms that drive this change, which include presence/absence of continental ice sheets, insolation, and complex ocean–atmosphere interactions. Our results indicate four distinct millennial-scale climate modes over the past 19 250 calendar years. These modes include a wet Last Glacial Maximum (19 250–17 120 cal. yr BP), a relatively dry late-Glacial/Holocene transition (17 120–9450 cal. yr BP), a wet early Holocene (9450–7670 cal. yr BP), and a highly variable mid-to-late Holocene climate (i.e., alternating wet/dry cycles; 7670 cal. yr BP–present). We attribute the mid-to-late Holocene climate interval to the onset of El Niño–Southern Oscillation ca. 7000 cal. yr BP and a more vigorous hydrologic system. These results are supported by a variety of regional terrestrial and marine palaeoclimate archives.

KEYWORDS: Lake Elsinore; sedimentology; southern California; climate change; ENSO.

Introduction

Hydrologic variability significantly impacts the environment and people of western North America (Wilkinson et al., 2002; Beuhler, 2003; Miller et al., 2003). For example, in California the droughts of 1976–77 and 1987–92 cost the state $2.6 billion (unadjusted) (Wilkinson et al., 2002). Southern California, home to more than 20 million people, has an arid, Mediterranean climate and faces a perennial freshwater shortage. As a result, this region is particularly susceptible to hydrologic variability over a range of temporal scales. To address this concern, it is essential to develop a baseline of past hydrologic variability. To date, there are relatively few published records of late Quaternary hydrologic variability for southern California. These records are limited to Mission diaries, tree-ring studies, some palynology from lagoons, and a few playa lake studies from the Mojave Desert (Lynch, 1931; Heusser, 1978; Meko et al., 1980; Davis, 1992; Enzel et al., 1992; Cole and Wahl, 2000; Biondi et al., 2001; D’Arrigo et al., 2001; Byrne et al., 2003). Kirby et al. (2004) recently published initial results from southern California’s largest, natural, non-playa lake: Lake Elsinore. These results indicate that Lake Elsinore contains exceptional potential for reconstructing southern California’s hydrologic variability over a variety of timescales.

As a part of the ongoing Lake Elsinore palaeoclimate study and its long-term objective of developing a natural baseline of climate variability, we present new multiproxy, lake sediment data that provide insight to millennial-scale hydrologic variability over the past 19 250 cal. yr BP. These new data indicate a series of a millennial-scale climate changes interpreted as variations in relative lake-level, or hydrologic balance. A combination of insolation, continental ice sheet dynamics, and complex ocean–atmosphere interactions are proposed as the primary forcing mechanisms of these millennial-scale changes in climate. Of interest, specifically, is a mid-Holocene climate shift to a more variable climate characterised by rapid hydrologic fluctuations. We interpret this shift as an indicator of the development of modern El Niño–Southern Oscillation
Study site

Lake Elsinore

Lake Elsinore is located 120 km southeast of Los Angeles, California and represents the largest natural, non-playa lake in coastal southwestern North America (Figs 1 and 2). The lake basin is a structural depression formed by offset motion along the Elsinore fault (Mann, 1956; Hull, 1991). Palaeoseismological research along a section of the Elsinore fault (~40 km south of Lake Elsinore) indicates that at least four major earthquakes have occurred over the past 4500 calendar years (Vaughan et al., 1999). The extent to which these earthquakes or the long-term evolution of the Elsinore basin impacts lake sedimentation over the past 19 250 calendar years is unknown. However, we are disinclined to attribute the observed changes in sedimentation over the past 19 250 calendar years to tectonics alone. Relatively constant sedimentation rates over the past 11 000 years from recently acquired drill cores from the lake’s deepest basin support our contention that tectonics, by itself, is not producing the observed sedimentological changes. Future seismic profiling of the lake basin will yield important insight into the basin’s tectonic–sedimentation relationship. The basin is estimated to contain > 600 m of sediment (Mann, 1956; Pacific Groundwater Digest, 1979; Damiata and Lee, 1986; Hull, 1991). Exploratory drill wells to > 500 m from the east end of the lake’s shore zone removed sediment described as mostly fine-grained (Pacific Groundwater Digest, 1979).

Lake Elsinore has a relatively small drainage basin (<1240 km²) from which the San Jacinto River flows (semi-annually) into and terminates within the lake basin (USGS, 1998). The San Jacinto River’s source is the San Jacinto Mountains, which lie 70 km to the northeast of Lake Elsinore and rise to a maximum height of 3290 m. An analysis of San Jacinto River discharge near San Jacinto, CA, over the interval AD 1920 to 2001 illustrates the river’s ephemeral behaviour with discharge limited to an annual average of 0.56 m³/s, 70% of which occurs during the months of February, March and April (Figs 1 and 3; (http://waterdata.usgs.gov)). In addition to the San Jacinto River, there are several smaller channels originating from the deeply incised, over-steepened Elsinore Mountains that confine the lake’s southwestern border (Fig. 2). These smaller channels indicate a likely sediment source during extreme precipitation events and/or wetter climate regimes.

Along the north shore of Lake Elsinore, there is a natural overflow channel, Walker Canyon. However, overflow through Walker Canyon has occurred only three times in the twentieth century and 20 times since AD 1769 (Lynch, 1931; USGS Lake-level Data, 2002). Based on the available data, each overflow event was very short-lived (< several weeks or months) demonstrating that Lake Elsinore is essentially a closed-basin lake system (Lynch, 1931; USGS Lake-level Data, 2002). Modern overflow elevation is 382.5 m, or approximately 5.5 m above the elevation of the cores discussed in this paper. Conversely, Lake Elsinore has desiccated on only 4 occasions since 1769 A.D. (Lynch, 1931; USGS Lake-level Data, 2002). The modern dry lake bed elevation is 371.2 m. Locals described the lake as ‘nothing more than a marshy patch of tules’ (perennial grasses) during periods of lake low stands (Mann, 1947). Sedimentological data from cores extracted from the deepest part of the profundal zone show no obvious evidence for periods of non-deposition during low stands (Kirby et al., 2004). As discussed in this paper, however, it is likely that cores from the littoral environment contain some intervals of non-deposition or sediment removal. Interestingly, 14C dated archaeological evidence from nearshore sediment deposits indicate that Lake Elsinore contained water nearly continuously over the past 8400 years, permitting humans to thrive permanently within the area since at least the early, mid-Holocene (Grenda, 1997). It is thought that any long-term lake drying or sustained flooding would have caused social upheaval and the subsequent abandonment of the nearshore occupation site.

Lake Elsinore is a polimictic lake with a 13 m maximum depth based on historic records (Anderson, 2001). Evaporation from the lake’s surface is estimated at > 1.4 m of water loss per year. From this estimate, the residence time of lake water is
Figure 2  Lake Elsinore aerial photograph (U.S. Department of the Interior). Inflow (San Jacinto River) and outflow (Temescal Wash) shown. Ephemeral washes associated Elsinore Mountains and sediment sources during large storms are shown as double-dashed white arrows.

Figure 3  San Jacinto River gauging station near San Jacinto, CA (http://waterdata.usgs.gov). Note the large range of flow variability. Y-axis extends to -0.5 to show better the years without data (e.g. 1992–1997).
probably very short, especially during drought periods (Mann, 1947; USGS, 1998; Anderson, 2001). There is a strong grain size gradient from the littoral zone (coarse-grained) to the profundal zone (fine-grained) attributed to wave action winnowing and resuspending the finer-grained component of the littoral sediments into the profundal environment (Mann, 1947). This process of littoral environment sediment reworking 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; Smoot, 2003; Smoot and Benson, 2004). Sediment trap studies indicate that CaCO3 is produced within the water column, probably linked to photosynthetic uptake of CO2 by phytoplankton (Anderson, 2001). SEM analyses of the lake sediment show distinct micro-metre size CaCO3 grains dispersed throughout the sediment (Anderson, 2001). There is also a strong relationship between mid-water sediment trap constituents and bottom water sediment trap constituents (Anderson, 2001). Sediment reworking from the littoral zone and resuspension from the profundal zone is an active process in Lake Elsinore; however, these processes are dependent on lake-level and wind speed (Anderson, 2001). There is no known source of detrital carbonate from within the lake’s drainage basin; although, the contribution of wind-blown, carbonate dust is unknown (Engel, 1959; McFadden et al., 1987; Reheis and Kihl, 1995). Lastly, the present-day surface distribution of CaCO3 is mirrored over time in the cores discussed in this paper. The highest CaCO3 values are located within the lake’s deepest basin and near the southeast edge (core 02-8; Anderson, 2001). The lowest values are found across the far southwest and northwest edges (cores 02-5 and 02-10; Anderson, 2001).

Regional climatology

The climate of coastal southwestern North America is Mediterranean, characterized by cool, ‘wet’ winters and hot, dry summers. Precipitation variability is dominated by the winter season (defined here as December–February) accounting for >50% (up to 60%) of the annual hydrologic budget; the inclusion of October, November, and March increase the annual contribution of winter precipitation to >80% (Lynch, 1931; USGS, 1998; Redmond and Koch, 1991; Friedman et al., 1992). As expected, the annual average precipitation for the region is positively correlated to winter average precipitation ($r=0.80; p<0.01$; Kirby, unpublished data), thereby illustrating winter season dominance of the annual precipitation cycle.

For coastal southwestern North America, winter precipitation amount is predominantly a function of storm trajectory as related to the winter season southward migration of the polar front (Weaver, 1962; Pyke, 1972; Lau, 1988; Schonher and Nicholson, 1989; Enzel et al., 1989, 1992; Redmond and Koch, 1991; Friedman et al., 1992; Ely, 1997). During the winter season, storms tracking across the region originate more frequently from the Pacific Ocean following a west/northwest-to-east/southeast storm path (Weaver, 1962; Pyke, 1972). Consequently, a copious precipitation falls along the predominantly northwest–southeast mountain ranges of coastal southwestern North America. This precipitation provides, generally, the necessary annual supply of freshwater recharge to the otherwise arid environment. The exact trajectory of storms over the study region is controlled largely by Pacific Ocean sea-surface conditions which modulate the overlying atmosphere and the average position of the winter season polar front (Namias, 1951; Weaver, 1962; Pyke, 1972; Namias and Cayan, 1981; Douglas et al., 1982; Lau, 1988; Namias et al., 1988; Schonher and Nicholson, 1989; Latif and Barnett, 1994; Trenberth and Hastenrath, 1994; Cayan et al., 1998; Dettinger et al., 1998).

Over the course of the winter season, the position of the polar front migrates both latitudinally and longitudinally. The extent of this migration, as previously explained, is strongly linked to sea-surface dynamics in the Pacific Ocean (Douglas et al., 1982; Namias et al., 1988; Schonher and Nicholson, 1989; Cayan et al., 1998; Dettinger et al., 1998). It is also well-documented that large-scale ocean–atmosphere linkages such as El Niño–Southern Oscillation and the Pacific Decadal Oscillation play a major role in the characteristics of winter season precipitation in coastal southwestern North America via their influence on polar front position and preferential storm paths (Lau, 1988; Schonher and Nicholson, 1989; Redmond and Koch, 1991; Ely et al., 1994; Cayan et al., 1998; Dettinger et al., 1998; Mantua, 2002). Observations indicate that years of strong El Niño activity generally increase the total winter precipitation in coastal southwestern North America (Redmond and Koch, 1991; Cayan et al., 1999; Castello and Shelton, 2004).

**Methods**

**Core collection**

Three sediment cores (LESS02-5 [91 cm], LESS02-8 [1302 cm], and LESS02-10 [688 cm]) were extracted from the littoral environment (approximately 376.7 m) during the 2002 lake-level low stand (Fig. 2). Core LESS02-5 was extracted using a manual push corer. Cores LESS02-8 and -10 were extracted using a hollow-stemmed auger drill core. Cores 02-5 and 02-8 were taken from within 10 m horizontal distance from one another at the east end of the lake. These cores are within 2 km of the pre-AD 1990 inlet of the San Jacinto River. Core LESS02-10 was taken from the west end of Lake Elsinore near the present-day city boat launch. Our assumption regarding sediment gaps is that all core drives using the hollow-stemmed auger drill core were to the measured depth. Any ‘missing’ core sections were subtracted from the top of the core’s individual drive and assumed to be a product of over-auguring between drives. Core LESS02-8 is missing approximately 6%; whereas core LESS02-10 is missing approximately 26%. We did not stack the cores to ‘fill-in’ sediment gaps knowing that cores from a littoral environment are subject to natural sediment hiatuses. Therefore, it is our opinion that a ‘stacked’ littoral core may in fact introduce a false sense of complete sediment recovery. Consequently, we present the core data as a series of individual drives per core site. Lastly, all cores were split, described, and archived in cold storage at Cal-State Fullerton.

**Magnetic susceptibility**

Samples were extracted from LESS02-5 and LESS02-10 at 0.5 cm over the first 100 cm; for the remainder of core LESS02-10 samples were extracted at 1.0-cm intervals. Core LESS02-8 was sampled at 1.0-cm intervals over its entire length. The samples were placed in preweighed 10-cc plastic cubes and magnetic susceptibility was measured twice on each sample with the y-axis rotated 180° once per analysis. All samples were analysed using a Bartington MS2 Magnetic...
Susceptibility instrument at 0.465 kHz. All magnetic susceptibility measurements were determined on the same day as cores were split and described to minimise possible magnetic mineral diagenesis with exposure to air. Following measurement of magnetic susceptibility, samples were reweighed to obtain total sediment wet 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 = χ) in SI units (×10⁻² m³ kg⁻¹).

LOI 550 °C (percentage total organic matter)

Total organic matter was determined using the loss on ignition method (Dean, 1974; Heiri et al., 2001). Samples were extracted from core LESS02-5 at 0.5-cm intervals and at 1.0-cm intervals in cores LESS02-8 and LESS02-10. All samples were dried at room temperature prior to grinding with a mortar and pestle. Ground samples were placed in a drying oven at 105 °C for 24 hours to remove excess moisture. Dried samples were transferred to preweighed crucibles, weighed to obtain dry sediment weight (average dry sediment weight = 4.67 g), and heated to 550 °C in an Isotemp® muffle oven for two hours. After two hours the samples were reweighed to obtain the percentage total organic matter from total weight loss.

LOI 950 °C (percentage total carbonate)

Total carbonate was determined also using the loss on ignition (LOI) method (Dean, 1974; Heiri et al., 2001). Samples were extracted from core LESS02-5 at 0.5-cm intervals and at 1.0-cm intervals in cores LESS02-8 and LESS02-10. Following the 550 °C analysis and weighing, crucibles were reheated to 950 °C for two hours in an Isotemp® muffle oven. After two hours the samples were reweighed and percentage total carbonate was calculated. As shown by Dean (1974), 3–4% total weight loss after 950 °C may be a function of clay dewatering. Consequently, we interpret values less than 3–4% as essentially zero total carbonate.

Results

Core descriptions

In the absence of quantitative grain size analysis, sediment descriptions are limited to subjective sand, silt, and clay classifications only. Colour changes and other notable materials are also shown (e.g. gastropods, organic layers, distinctive lithologic components). Core LESS02-5 is described elsewhere in detail by Kirby et al. (2004). There are several features worth noting in cores LESS02-8 and -10 (Figs 4 and 5). Both cores are clay-rich from the bottom to 560 cm in 02-8 and 587 cm in 02-10. The bottom clay unit is also characterised by distinct CaCO₃ flecks in 02-8. Above this clay unit, the sediment becomes silty and grades into a carbonate-rich interval characterised by CaCO₃ flecks and some rootlets. At 301 cm and 365 cm in 02-8 and 02-10, respectively, the sediment becomes notably darker and coarser. From the latter depths to the core tops, the sediment contains intervals of gastropods, clay pock- ets, rootlets and organic-rich horizons.

Magnetic susceptibility

Magnetic susceptibility values from core LESS02-8 have a maximum value of 2.8 × 10⁻² m³ kg⁻¹ and a minimum value of 0.7 × 10⁻² m³ kg⁻¹ (Fig. 4). Values from core LESS02-10 range from a maximum of 6.1 × 10⁻² m³ kg⁻¹ to a minimum of 0.8 × 10⁻² m³ kg⁻¹ (Fig. 5). Both cores can be divided into three sections on the basis of magnetic susceptibility. The bottom section (688 to 425 cm in core 8 and 684 to 495 cm in core 10) represents the lower portion of both cores where the magnetic susceptibility values are low, but variable (Figs 4 and 5). The middle section (425 to 305 cm in core 8 and 495 to 365 cm in core 10) represents the central portion of both cores where the magnetic susceptibility values are uniformly low (Figs 4 and 5). The upper section (305 to 0 cm in core 8 and 365 and 0 cm in core 10) is the top portion of both cores where the magnetic susceptibility values are high and variable.

Age control

In addition to the dates described in Kirby et al. (2004) from core LESS02-5, seven additional dates were obtained from core LESS02-8 and 02-10 at locations demarcating distinct sedimentological change (Table 1; Figs 4 and 5). All dates were measured at BETA Analytic, Inc. or the University of California, Irvine Keck AMS Facility, using bulk sediment organic carbon treated with an acid wash to remove carbonate. The lack of salvageable macro- or micro-organic matter (rootlets: see Figs 4 and 5) precluded the dating of discrete organic detritus. As shown by Kirby et al. (2004) surface sediments from the lake’s deepest basin provide a modern date; as a result, we do not consider our bulk sediment organic carbon dates to be affected significantly by old carbon.

Table 1 Radiocarbon dates

<table>
<thead>
<tr>
<th>Identification</th>
<th>Core Depth (cm)</th>
<th>Conventional Radiocarbon Age</th>
<th>Calibrated Age* (midpoints cal. yrs. B.P.)</th>
<th>2-Sigma Calibrated Age* (cal. yrs. B.P.)</th>
<th>δ¹³C</th>
</tr>
</thead>
<tbody>
<tr>
<td>Beta-169437</td>
<td>LESS02-5 (21–22 cm)</td>
<td>2010 ± 40 BP</td>
<td>1960 cal yrs. B.P.</td>
<td>2050 to 1880</td>
<td>−23.1</td>
</tr>
<tr>
<td>Beta-169438</td>
<td>LESS02-5 (78–79 cm)</td>
<td>3220 ± 40 BP</td>
<td>3450 cal yrs. B.P.</td>
<td>3470 to 3390</td>
<td>−23.3</td>
</tr>
<tr>
<td>UCAMS-10336</td>
<td>LESS02-8 (141–143 cm)</td>
<td>6845 ± 25 BP</td>
<td>7670 cal yrs. B.P.</td>
<td>7610 to 7740</td>
<td>−22.4</td>
</tr>
<tr>
<td>UCAMS-10337</td>
<td>LESS02-8 (300–302 cm)</td>
<td>8390 ± 40 BP</td>
<td>9450 cal yrs. B.P.</td>
<td>9400 to 9490</td>
<td>−23.3</td>
</tr>
<tr>
<td>UCAMS-10338</td>
<td>LESS02-8 (462–463 cm)</td>
<td>12570 ± 60 BP</td>
<td>14900 cal yrs. B.P.</td>
<td>14260 to 15540</td>
<td>−32.2</td>
</tr>
<tr>
<td>UCAMS-10340</td>
<td>LESS02-8 (650–651 cm)</td>
<td>15260 ± 60 BP</td>
<td>18270 cal yrs. B.P.</td>
<td>17700 to 18830</td>
<td>−26.7</td>
</tr>
<tr>
<td>UCAMS-10341</td>
<td>LESS02-8 (685–686 cm)</td>
<td>16120 ± 70 BP</td>
<td>19250 cal yrs. B.P.</td>
<td>18640 to 19860</td>
<td>−24.7</td>
</tr>
<tr>
<td>Beta-179161</td>
<td>LESS02-10 (308–310 cm)</td>
<td>6570 ± 40 BP</td>
<td>7450 cal yrs. B.P.</td>
<td>7560 to 7420</td>
<td>−24.9</td>
</tr>
<tr>
<td>Beta-185174</td>
<td>LESS02-10 (494–496 cm)</td>
<td>14250 ± 50 BP</td>
<td>17120 cal yrs. B.P.</td>
<td>17470 to 16770</td>
<td>−22.6</td>
</tr>
</tbody>
</table>

*Stuiver et al., 1998.
Precaution was taken not to select sediment for dating from intervals interpreted as low stands to avoid obtaining false ages via reworked older carbon (Kirby et al., 2004; Smoot and Benson, 2004). Dates were cross-correlated among cores using distinct high and low values on the magnetic susceptibility curves. Figure 6 illustrates the points of cross-correlation among cores LESS02-5, LESS02-10, and LESS02-8 using magnetic susceptibility. The correlation between LESS02-5 and LESS02-10 has been previously used for cross-dating (Kirby et al., 2004). Figure 6 demonstrates clearly the ability to transfer dates basin-wide using core sediment data for Lake Elsinore. For the purpose of this research, nine dates over 7 m provide sufficient age control for the assessment of millennial-scale climate interpretations.

We did not construct an age model because of possible unrecognised hiatuses associated with littoral environment...
sedimentation. Similar to our decision not to ‘stack’ the cores, we feel that an age model for our littoral cores is specious without full knowledge of the likely gaps in sedimentation. However, for the purposes of characterising millennial-scale hydrologic variability over the past 19,250 years, our distribution of dates is satisfactory. As a result, we limit our interpretations to first-order climate interpretations only.

**Figure 5** LESS02-10 core lithology, magnetic susceptibility, and dates (cal. yr BP). Key provided for identification of lithologic features. No recovery intervals are labelled. Correlated dates are underlined; dates from core LESS02-8 are highlighted.

Total organic matter (TOM) in core LESS02-8 ranges from 5.9 to 0.9% (Fig. 7). In core LESS02-10, these data range from 4.3% to 0.7% (Fig. 8). Unlike the magnetic susceptibility data, the total organic matter is less well-defined by ‘sections’. In core 8, percentage total organic matter is highest.

LOI 550°C (total organic matter)
near the core bottom, dramatically decreases from 575 to 550 cm, and then remains low until 445 cm (Fig. 7). Values rise again at 445 and remain high until 390 cm, at which point they decrease gradually to the core top. The upper 225 cm of core 8 is also characterised by several abrupt, high-amplitude changes. The profile of %TOM in core 10 is very similar to core 8 except for an interval of high values between 280 and 310 cm that is not present in core 8 (Fig. 8).

LOI 950 °C (total carbonate)

Total carbonate (TC) in both core LESS02-8 and LESS02-10 ranges from ~20% to <3%, or zero according to Dean’s (1974) research on the limitations of LOI analysis (Figs 6 and 7). Core 8 is characterised by low, variable TC values from the core bottom to 565 cm. From 565 to 505 cm, TC values are uniformly low. At 505 cm, the values abruptly rise to 12% and remain high and variable to 390 cm, where they decrease and remain at ~6% to 235 cm. From 235 to 190 cm, the values again increase with large-amplitude variability. An abrupt decrease occurs at 190 cm and lasts until 145 cm. From 145 to 15 cm, %TC is relatively high and uniform. Values again decrease at 15 cm though to the core top. Core 10%TC data are much less variable (see ‘Discussion’ for explanation) than core 8. A notable exception is the interval between 490 and 365 cm. Here the values are initially high and variable and monotonically decrease to low values by 365 cm. A brief excursion to higher values occurs at 230 cm core depth.

We note that there are no significant correlations between magnetic susceptibility, LOI 550 °C, or LOI 950 °C over the length of the records for either core.

Discussion

Magnetic susceptibility

Sediment magnetic susceptibility measures the concentration of magnetic material in a sediment sample (Thompson et al., 1975). In clastic-dominated sediment systems such as Lake Elsinore, the amount of magnetic material is a function largely of the flux of allochthonous sediment into the lake’s basin. There are, however, other factors that can affect sediment magnetic susceptibility such as post-depositional diagenesis (Hilton
and Lishman, 1985; Hilton et al., 1986), dilution from organic
matter and/or carbonates, and mineral segregation by wave
action (Smoot and Benson, 2004). Because both cores
LESS02-8 and -10 were collected from near the modern shore
position, it is expected that the sediments are influenced
strongly by wave action and similar nearshore current pro-
cesses throughout similar lake stands (Fig. 2). As shown by
Mann (1947) and Anderson (2001), a strong grain size gradient
exists between the littoral environment (coarser grains) and the
profundal environment (finer sediments). Finer-grained mag-
netic minerals are more susceptible to diagenetic alteration
(Maher and Thompson, 1999). As a result, variations in mag-
netic susceptibility in the coarser-grained, littoral sediments
used for this study are not likely to have been caused by diage-
netic alteration. Instead, we hypothesise that the waxing and
waning of lake-level—a function of climate—and the episodic
flux of clastics—a function of extreme precipitation events—
are the dominant controls on the magnetic susceptibility values
in cores LESS02-8 and 02-10 (Figs 4, 5, 7 and 8). The absence of
a correlation between magnetic susceptibility and LOI 550 °C or
LOI 950 °C supports our contention that the magnetic sus-
ceptibility records changes in clastic flux and not dilution via
organics or carbonates.

In terms of a climatic explanation, we contend that a wetter
climate produces higher magnetic susceptibility values for two
reasons: (1) a wetter climate raises lake-level, placing the core
location under more water and reducing the removal of mag-
netic minerals by wave action winnowing; and (2) an increase
in the frequency of extreme precipitation events combined
with a generally wetter climate increases the flux of mag-
netic-rich sediments to the littoral environment. Research by
Kirby et al. (2004) supports this hypothesis by observing a posi-
tive correlation between twentieth-century lake-level variabil-
ity and magnetic susceptibility in Lake Elsinore (i.e. higher
lake-levels = higher magnetic susceptibility). Inman and
Jenkins (1999) have shown a strong positive correlation over
the twentieth century between climate wetness and river sedi-
ment flux in southern California further supporting a climate–
sediment flux hypothesis. Using the results of Kirby et al.
(2004), we extend, in this paper, our interpretation of magnetic
susceptibility as a proxy for hydrologic balance (i.e. lake-level)
over the past 19 250 calendar years. It is noted that several
authors find a similar relationship between lake-level and mag-
netic susceptibility; however, the underlying control on this
relationship may vary from basin to basin (e.g. Benson et al.,
1998; Negrini et al., 2000).

LOI 550 °C (total organic matter)

Total organic matter in lake systems reflects a combination
of autochthonous and allochthonous sources. Allochthonous
organic matter is derived from terrestrial organic matter washed
into the lake basin. The extent to which this fraction contributes
to lake sediments is controlled by several factors, including

Figure 7  LESS02-8 sedimentology, dates, and climatic interpretations. Left: mass magnetic susceptibility; middle: LOI 550 °C (percentage total
organic matter); right: LOI 950 °C (percentage total carbonate). Dates in calendar years before present are shown adjacent to the depth scale. Bold
dates are from core LESS02-8; italicised dates are cross-correlated from either core LESS02-5 or 02-10.
Drainage basin flora, regional climate, and basin topography (Meyers and Ishiwatari, 1993). In most lakes, except for extremely oligotrophic lakes, the autochthonous source dominates (Dean and Gorham, 1998). Autochthonous organic matter derives from in situ lake productivity of various types (Meyers and Ishiwatari, 1993). Anderson (2001) has shown that modern profundal sediments in Lake Elsinore average 4.84% organic carbon; modern littoral sediments contain 0.79% organic carbon. In both environments, carbon:nitrogen data show that autochthonous lake organics dominate. The gradation in total organic carbon from littoral to profundal environment is produced by nearshore wave action winnowing that removes preferentially the finer-grained organic matter and transports it into the deeper basins.

The response, however, of sediment organic matter to lake-level change is less straightforward. Both cores show their highest total organic matter values during the Last Glacial Maximum (LGM) (Figs 7 and 8). Using this first-order observation, we suggest that the higher values during the LGM reflect a more diverse and substantial regional flora associated with cooler climates (Wells and Berger, 1967; Spaulding, 1990; Wells et al., 2003). This more diverse and substantial flora produced a net increase in the total allochthonous organic matter washed into the lake, combined with the ambient autochthonous signal, thus increasing total sediment organic matter. Future geochemical analyses will help to evaluate the veracity of the latter statement.

Crop 950°C (total carbonate)

In many lake systems, the carbonate sediment fraction is produced within the lake through a variety of possible processes (Thompson et al., 1997; Mullins, 1998; Hodell et al., 1998; Benson et al., 2002; Kirby et al., 2002). Because there is no local source of detrital carbonate within Lake Elsinore's drainage basin, it is assumed that the carbonate in the lake's sediment is entirely autochthonous; although the influence of wind-blown carbonate dust cannot be ruled out at this point (Reheis and Kihl, 1995). In addition, we cannot rule out mixing of waters with different source areas as a cause of changes in the relative saturation of lake waters with respect to calcium carbonate. Future δ18O(calcite) analyses on the calcite fraction may help to resolve these issues. Anderson (2001) has shown that carbonate precipitates directly within the water column in response to CO2 drawdown by phytoplankton. There is also a strong seasonal contrast in carbonate precipitation that indicates a temperature role. Hydrologic models by Anderson (2001) show that carbonate precipitation will increase as lake-levels decrease in response to saturation of the water column with respect to calcium and carbonate. This finding agrees with our research, which shows that the highest percentage of carbonate occurs during inferred low lake-levels (Figs 4, 5, 7 and 8). Lithologically, these low stands are evident by the abundance of large carbonate nodules, gastropods, rootlets and organic layers (Figs 4 and 5). From this observation, we
interpret total carbonate, like magnetic susceptibility, as a first-order proxy for relative lake-level. Of the three proxies discussed, only total carbonate shows dramatic differences between cores. The explanation for this intra-basin disparity is not clear. However, modern analyses of surface CaCO₃ distribution show similar disparities between the two core sites both in absolute magnitude of differences and site location (Anderson, 2001).

Millennial-scale Holocene hydrologic variability at Lake Elsinore

Over millennial timescales, hydrologic variability in southern California is controlled by first-order forcings such as insolation, glacial boundary conditions (i.e. presence/absence of continental ice sheets), and complex ocean-atmosphere interactions. Over the past 19 250 calendar years, winter and summer insolation at 30°N latitude has changed dramatically (Fig. 9). These changes affect the strength of seasonality, the mean position of the polar front jet stream (Kirby et al., 2002), and its associated storm tracks (e.g. Sawada et al., 2004). The extent and morphology of the North American ice sheet also impacts climate through its modulation of atmospheric circulation. In fact, Negrini's (2002) summary of Great Basin lake histories during the Quaternary attributes the first-order patterns of lake-level change to the mean latitude of the polar front jet stream as controlled by a combination of insolation and glacial boundary conditions. A lower latitude polar front favours the advection of moisture-rich storms across western North America. Of course, as the ice sheet decays, its influence on the position of the jet stream decreases. Moreover, as the ice sheet decays the jet stream will migrate north, thus imprinting a spatial progression of lake-level change from south to north (Cole and Wahl, 2000; Enzel et al., 2003). With
the disintegration of the North American ice sheets, the relative impact of the ice sheet on climate, specifically atmospheric circulation and storm tracks, will cease. It is also suggested that other, less understood forcing mechanisms may contribute to Holocene climate change in the region. Additional climate forcings on millennial timescales include solar variability and complex ocean–atmosphere interactions (e.g. Bond et al., 2001; Friddell et al., 2003).

We suggest that Lake Elsinore's inferred climate history reflects a combination of insolation, ice sheet, and complex ocean–atmosphere climate forcings (Fig. 9). Figure 9 summarises the first-order, inferred lake-level history for Lake Elsinore with its associated climate forcings. From the base of the record (19 250 cal. yr BP) to 17 120 cal. yr BP (LGM), lake-levels are interpreted as relatively high and stable at the core location. Low carbonate values during this interval suggest that lake-levels were above the core site (Figs 7 and 8). Total organic matter is initially high in both cores during this high stand. These high organic values are interpreted as evidence for a greater influx of terrigenous organic carbon via runoff into the lake basin as well as a more diverse and extensive terrestrial biomass (Figs 7 and 8). During the latter half of this high stand, total organic matter values decrease. Rapid throughput of lake water during this peak high stand may have favoured an oligotrophic lake as nutrients passed through the system rapidly.

Magnetic susceptibility values are low, but variable throughout this interval. As suggested previously by Kirby et al. (2004), low magnetic susceptibility values are interpreted as evidence for low lake-levels in Lake Elsinore. Yet, here we suggest that lake-levels were high. We propose two explanations for this apparent contradiction between low magnetic susceptibility values and our interpretation of high lake-levels during the LGM interval. First, the magnetic minerals may have been diluted by an increase in organic matter (Figs 7 and 8). The absence of a correlation between magnetic susceptibility and organic matter through this interval weakens a dilution explanation. Second, the source of magnetic minerals during the LGM may be different from the Holocene's source. We favour the latter explanation based on the general relationship between the occurrence of coarse-grained sediment and magnetic susceptibility values. A comparison of core lithology to magnetic susceptibility shows a strong first-order relationship between the occurrence of coarse-grained sediment and high magnetic susceptibility values (Figs 4 and 5). The occurrence of coarse-grained sediment (i.e. gravel) does not appear in significant quantities until 300 and 355 cm (9450 cal. yr BP) in cores LESS02-8 and 02-10, respectively. In both cores, magnetic susceptibility values rise abruptly in tandem with this increase in coarse-grained sediment. Sediment deposited during the LGM interval are generally finer-grained and devoid of coarse gravel (Figs 4 and 5). We suggest that this change in sediment size and its associated magnetic susceptibility values reflects a switch in sediment provenance between the LGM and the Holocene. During the LGM, the San Jacinto River was less ephemeral than during the Holocene as the length of the melt season from the high-elevation San Jacinto Mountains lingered into the summer (perhaps autumn?). We propose that the sediment derived from the distal San Jacinto Mountains via the San Jacinto River during the LGM was finer-grained (due to a longer transport path or increased weathering due to a wetter climate?) and lacked the larger sediment particles that raise magnetic susceptibility values. Alternatively, or perhaps in conjunction with the latter hypothesis, the San Jacinto Mountains represent a lower magnetic susceptibility provenance. As climate changed and the influence of the San Jacinto River sediments diminished, the effect of local sediment sources (i.e. Elsinore Mountains) increased. The shorter transport path from the proximal Elsinore Mountains is reflected in the coarser-grained sediments of Holocene age. Therefore, we interpret the proposed high lake-levels during the LGM, despite low magnetic susceptibility values, as evidence for a change in the relative influence of the San Jacinto River and sediment transport distance. The recent October 2004 storms in southern California raised Lake Elsinore lake-level by 0.56 m in two weeks (P. Kilroy personal communication, 2004). This modern system response indicates the importance of the San Jacinto River to maintaining high lake-levels in Lake Elsinore as well as the lake's sensitivity to large storms. Using this modern example, we feel that higher lake-levels during the LGM must be related to a more 'regular' flow of the San Jacinto River in response to more frequent, large storm events.

This LGM relative high stand was forced by the presence of an ice sheet over northern North America combined with relatively cool summers (Fig. 9). The ice sheet depressed the mean annual latitude of the polar front, increasing the frequency of storms across the study area (Fig. 9). Combined with cooler summers, Lake Elsinore persisted without large seasonal lake-level variations. Our interpretation is supported by pluvial lake records from the Mojave Desert and other regions of southwestern North America that show similar high and stable lake-levels during the LGM (Enzel et al., 2003; Wells et al., 2003). Paaleoecological records also indicate a wetter Glacial Maximum climate for this region (Wells and Berger, 1967; Spaulding, 1990; Mock and Bartlein, 1995).

Lake-levels decreased following the LGM wet climate interval (Figs 7–9). This low stand occurred during the late-Glacial/ Holocene transition (17 120 to 9450 cal. yr BP). An abrupt rise in total carbonate, including the occurrence of large, in situ carbonate nodules, is interpreted as evidence for low lake-levels. The supersaturation of shallow water and/or porewaters with Ca$^{2+}$, HCO$_3^-$, and CO$_3^{2-}$-favoured copious carbonate production. As noted by Rosen (1994), the occurrence of evaporites, particularly 'displacive evaporites' (i.e. carbonate nodules), is strong evidence for the presence of water, albeit shallow, intermittent, or solely within near-surface sediment pores. Total organic matter values are variable, but generally higher than average during this interval. Here, the higher-than-average total organic values are attributed to the occurrence of shallow-water macrophytes and/or the periodic growth of near-shore grasses. This interpretation is supported by the observation of rootlets and occasional organic-rich layers (Figs 4 and 5). In the modern system, low lake-levels are characterised by the rapid proliferation of grasses along the shore environment (Mann, 1947). Magnetic susceptibility values are extremely low and uniform during this interval. Low, uniform magnetic susceptibility values are attributed to a reduction of eroded magnetic-rich minerals from the surrounding terrain in response to decreased precipitation. Abrupt retreat of the North American ice sheet in conjunction with greater seasonality, especially warmer summers, favoured both a decrease in storm frequency across the study region and more intense summer heatwaves. Together, these millennial-scale climate forcings favoured a prolonged lake-level low stand. Our interpretation of the late-Glacial/Holocene transition as a dry interval is supported by the occurrence of intermittent, shallow lakes in the Mojave Desert (Wells et al., 2003).

Following the late-Glacial/Holocene transition low stand, there was a shift to a wetter climate. This climate shift occurred in the early Holocene, dated from 9450 cal. yr BP to 7670 cal. yr BP (Figs 7–9). We distinguish this interval because of its position separating the well-defined late-Glacial/ Holocene transition and the mid-to-late Holocene climate interval. Total organic matter and total carbonate are highly
variable; although, the data indicate at least one period of low lake-levels as recognised by higher CaCO$_3$ values in both cores and the occurrence of CaCO$_3$ flecks in core 02-8 (Figs 4, 5, 7 and 8). Magnetic susceptibility values show a general increase over the interval (Fig. 4). This coincides with the first regular occurrence of coarse-grained sediment (Figs 4 and 5). We interpret the early Holocene interval as a time of rising lake-levels in response to an increase in summer precipitation (Figs 7–9). Maximum seasonality during the early Holocene would have enhanced the North America monsoon as well as the occurrence of local convection during the late summer and early autumn months (Wells and Berger, 1967; Spaulding, 1990). At the same time, winter insolation values were at a minimum and may have increased the frequency of winter storms. The combination of summer monsoonal rains and more frequent winter storms may have increased the erosion of local sediment sources such as the Elsinore Mountains. As a result, the occurrence of large coarse-grained sediments increases, which produced higher magnetic susceptibility values (Figs 4 and 5). Packrat midden studies from the Mojave and southern Great Basin indicate a wet early Holocene climate attributed to the influence of an enhanced early Holocene North American monsoon (Wells and Berger, 1967; Spaulding, 1990). Similarly, sediments from Dry Lake in the San Bernardino Mountains reveal large storm deposits in the early Holocene that may be linked to enhanced summer storm activity.

After the early Holocene climate shift came the mid-to-late Holocene climate interval from 7670 cal. yr BP to the present (Figs 7–9). Sediment deposited during this interval is characterised by variable total organic values. Total carbonate is also characterised by greater variability, except in core LESS02-10 where its present-day location is characterised by low surface carbonate values (Anderson, 2001). Magnetic susceptibility data are among the highest on record, and they, too, are characterised by significant variability. The occurrence of coarse-grained sediment also increases during this interval (Figs 4 and 5). Again, there are no significant correlations between magnetic susceptibility and LOI 550°C or LOI 950°C during the mid-to-late Holocene interval. Based on these data, we interpret this interval as a period of highly variable climate alternating between extreme wet and dry intervals. The observation that ~7000 years is represented by only 1.5 m of sediment is strong evidence for significant depositional hiatuses or sediment removal during the mid-to-late Holocene. One of the most obvious ways to explain these abrupt changes in sedimentation is to have a rapidly rising and falling lake-level, which impedes the accumulation of thick, undisturbed sediment packages. There are multiple lines of evidence to support our interpretation of a highly dynamic mid-to-late Holocene climate. In the Mojave Desert, sedimentological evidence suggests the occurrence of brief lakes throughout the mid-to-late Holocene (Enzel et al., 1989; Enzel, 1992; Enzel et al., 1992; Enzel and Wells, 1997; Wells et al., 2003). Palynological evidence from the San Joaquin Marsh (Orange County, CA) shows distinct periods of increased flooding and erosion during the late-Holocene (Davis, 1992). Davis (1999b) also uses pollen to infer intervals of a wetter mid-to-late Holocene in Tulare Lake, approximately 250 km northwest of Los Angeles. Cole and Wahl (2000) use a combination of sedimentological and palynological data from a lagoon near San Diego to infer a wetter climate during the late-Holocene. From Lake Elsinore, itself, Kirby et al. (2004) show a distinct oscillation between high and low lake-levels over the last 3800 yr BP. From the marine realm, $^{18}$O values from diatoms in the Gulf of California show strong climate variability over the late-Holocene (Jullet-Leclerc and Schrader, 1987). Pollen records from the Santa Barbara basin indicate a rise in total pollen after 5000 $^{14}$C yr BP perhaps in response to greater erosion (Heusser, 1978). Also from the Santa Barbara basin, Fridell et al. (2003) infer an increase in climate variability beginning in the mid-Holocene based on $^{18}$O values from foraminifera. And, most recently, Barron et al. (2004) show an increase in surface productivity in the Gulf of California, which they interpret as evidence for more frequent El Niño forced wind excursions. Together, these data, both terrestrial and marine, support a more variable mid-to-late Holocene climate.

Evolution of modern El Niño–Southern Oscillation

Why did climate variability increase in the mid-to-late Holocene? The relationship between hydrologic extremes and hydrologic variability in southern California to El Niño–Southern Oscillation is well documented (Schonher and Nicholson, 1989; Redmond and Koch, 1991; Piechota et al., 1997; Cayan et al., 1999). Although the forcing mechanism driving ENSO remains debated, it is clear that changes in sea surface thermal structure, the propagation of ocean thermal anomalies, and their effect on atmospheric circulation result in hydrologic variability along the west coast of North America (Latif and Barnett, 1994; Gu and Philander, 1997; Pierce, 2002). Generally, El Niño years produce more precipitation in southern California and La Niña years produce less (Redmond and Koch, 1991; Castello and Shelton, 2004). In addition, ENSO-type interannual climate variability is superimposed on decadal patterns of climate change that may reflect ‘sustained’ intervals of a preferred climate mode (Douglas et al., 1982; Jacobs et al., 1994; Trenberth and Hurrell, 1994; Cayan et al., 1998; Toure et al., 2001; Mokhov et al., 2004). From these analyses, it is reasonable to suggest that the conditions that produce modern ENSO are dependent on specific forcing mechanisms that evolve through time (e.g. the presence of continental ice sheets in response to Milankovitch forcing; Loubere et al., 2003). Research indicates that ENSO exists under a variety of climate boundary conditions; although, the amplitude, duration, and frequency is subject to change (Clement et al., 2000; Tudhope et al., 2001; Moy et al., 2002). Owing to the global effect of ENSO, there has been considerable research on the evolution of modern ENSO conditions.

Research from South America, where ENSO is first detected, indicates that modern ENSO conditions evolved into the present state over the Holocene (Sandweiss et al., 1996; Rodbell et al., 1999; Sandweiss et al., 2001; Andrus et al., 2002; Moy et al., 2002). Although the exact timing is still debated, it is clear that this transition occurred by the mid-Holocene (~5000 cal. yr BP). Recently, Fridell et al. (2003) published $^{18}$O$_{calc}$(calcite) data from foraminifera in the Santa Barbara Basin. They conclude that climate variability, perhaps linked to more intense El Niño events, increased by the mid-Holocene. To the north, Barron et al. (2003) reach a similar conclusion, although slightly later in the Holocene (~3500 cal. yr BP), using multi-proxy data from ODP Site 1019 adjacent to northern California. And, to the south, Barron et al. (2004) suggest an increase in ENSO variability beginning at 6.2 ka. In the context of these studies, we also interpret our data in terms of ENSO evolution. As previously stated, we interpret the mid-to-late Holocene climate interval at Lake Elsinore as highly dynamic and variable. This interpretation is congruent with the evolution of ENSO and its known impact on southern California hydrology. The exact forcing mechanism driving the Holocene evolution of ENSO is unclear; however, at millennial timescales, it is probably related to long-term changes in the
seasonal cycle of insolation and its effect on latitudinal thermal gradients (Clement et al., 2000).

Conclusion

New results from southern California’s largest, non-playa lake, Lake Elsinore, reveal millennial-scale patterns of climate change over the past 19 250 calendar years. A wet climate during the LGM (19 250–17,120 cal. yr BP) is attributed to the presence of the North American ice sheet and its southward deflection of the polar front jet stream. Subsequent ice sheet retreat and decay coincided with warming summers through changes in insolation produced a dry late-Glacial/Holocene transition (17 120–9450 cal. yr BP). During the early Holocene (9450–7670 cal. yr BP), the contribution of summer precipitation increased in response to maximum insolation-produced seasonality and its strengthening of the North American monsoon. The mid-to-late Holocene (7670 cal. yr BP–present) was characterised by a more vigorous hydrologic system, which produced a highly variable climate and the rapid lake-level fluctuations at Lake Elsinore. A more vigorous hydrologic system is attributed to the onset of modern ENSO conditions. These conditions increased the frequency of large winter storms across the study region. Regional palaeoclimate archives from a combination of marine and terrestrial locations support our interpretation of both millennial-scale climate change over the last 19 250 cal. yr BP and our contention that ENSO plays a role in mid-to-late Holocene hydrologic dynamics in southern California. Although low-resolution, this research provides important regional data for a more complete understanding of the spatial and temporal climate dynamics in southern California over the past 19 250 calendar years.

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