1997

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Benson, Larry; Burdett, James; Lund, Steve; Kashgarian, Michaele; and Mensing, Scott, "Nearly synchronous climate change in theNorthern Hemisphere during the last glacial termination" (1997). *USGS Staff -- Published Research*. 793.
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Nearly synchronous climate change in the Northern Hemisphere during the last glacial termination

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The climate of the North Atlantic region underwent a series of abrupt cold/warm oscillations when the ice sheets of the Northern Hemisphere retreated during the last glacial termination (17.7–11.5 kyr ago). Evidence for these oscillations, which are recorded in European terrestrial sediments as the Oldest Dryas/Bølling/Older Dryas/Allerød/Younger Dryas vegetational sequence[1,2], has been found in Greenland ice cores[3,4]. The geographical extent of many of these oscillations is not well known[5,6], but the last major cold event (the Younger Dryas) seems to have been global in extent[7–10]. Here we present evidence of four major oscillations in the hydrological balance of the Owens basin, California, that occurred during the last glacial termination. Dry events in western North America occurred at approximately the same time as cold events recorded in Greenland ice, with transitions between climate regimes in the two regions taking place within a few hundred years of each other. Our observations thus support recent climate simulations which indicate that cooling of the North Atlantic Ocean results in cooling of the North Pacific Ocean[11] which, in turn, leads to a drier climate in western North America[12].

Owens Lake is located in the Great Basin of the western United States between the central Sierra Nevada and the Inyo-White mountains. Maximum precipitation along the Sierra Nevada is associated with the annual north–south progression of the polar jet stream, with cool-season orographic precipitation from North Pacific sources supplying >99% of the runoff reaching the Owens basin[13,14].

Core OL84B was obtained from the Owens basin in 1984 using a modified Livingstone piston corer (Fig. 1)[15]. A series of accelerator mass spectrometry (AMS) ^{14}C dates on bulk organic carbon were used to set the chronology of the core section used in this study (Fig. 2). The ^{14}C-based records of Owens Lake were converted to calendar years using ^{230}Th–^{234}U and ^{14}C ages of corals[16–19]. The ^{14}C data indicate sediment hiatuses at 2.25 and 9.20 m. The upper surface of the 9.20-m hiatus contains features indicating desiccation and subaerial exposure, including a lag deposit (1–3 mm thick) of frosted quartz grains. The 2.25-m hiatus is marked by a coarse sand. Sediments examined in this study are thus bounded by two desiccation events that occurred ~18.3–15.8 and ~6.7–4.5 kyr ago (Fig. 3)[20]. The younger hiatus, previously noted in a core taken from higher elevation in the Owens basin[21] seems to have occurred at about the same time as hiatuses noted in cores from the Mono Lake basin[22]. The older hiatus appears to coincide with major declines in the levels of Mono Lake and Lake Lahontan[23,24].

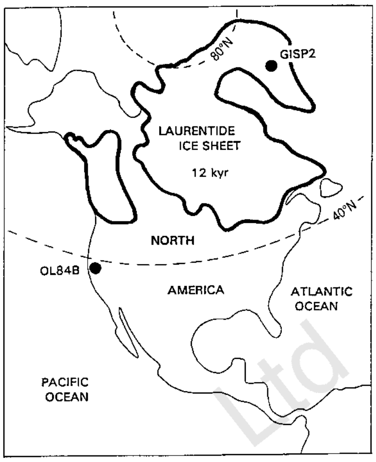

Figure 1 Map of North America showing locations of sediment core OL84B and ice core GISP2.

To determine the degree of variability in the hydrological balance of Owens Lake between 15.8 and 6.7 kyr, we analysed a continuous set of 120 samples for δ^{18}O and total inorganic carbon (TIC) (Fig. 3). Each sample integrated ~60 yr of time and was repeatedly washed with distilled–deionized water to remove soluble salts. The isotopic analyses were done on the TIC fraction.

In lakes that oscillate between closed and open hydrological states, δ^{18}O and TIC values tend to decrease with decreasing residence time of water in the lake basin[20]. The δ^{18}O value of a lake is a function of the overflow : inflow ratio. When this ratio approaches unity, the δ^{18}O value of a lake approaches the δ^{18}O value of inflow; when this ratio approaches zero, the δ^{18}O value of lake water becomes highly enriched owing to evaporative concentration of the heavy isotope of oxygen (^{18}O) in lake water. For Owens Lake, calcites precipitated during high overflow : inflow conditions would have δ^{18}O values approaching ~15‰ (relative to the VSMOW standard); in contrast, calcites precipitated during hydrological closure would have δ^{18}O values approaching ~30‰ (ref. 20).

During overflow, losses of Ca^{2+} (and CO_3^{2-}) increase with increase in the overflow : inflow ratio; therefore, if the flux of detrital silicates remains roughly constant, the TIC fraction becomes a proxy for change in hydrological balance. During hydrological closure, all Ca^{2+} that reaches a lake eventually precipitates as $CaCO_3$. The TIC fraction in a closed lake tends to decrease with increasing lake size because the flux of suspended sediment increases exponentially with increasing discharge to the lake, diluting the carbonate precipitate[25]. Glaciation of a lake basin's catchment area also can markedly increase the flux of detrital silicates to a lake, completely masking the TIC signal[20].

The δ^{18}O data obtained in this study indicate four extremely dry (closed-basin) intervals (D_1 to D_4) that occurred between 15.8 and 6.7 kyr (Fig. 3). The older three dry intervals are centred at 15.1, 13.2 and 12.2 kyr ago; the youngest dry interval (D_4) marked the beginning of the Holocene (11.3 kyr) and extends to Desiccation II. The presence of prismatic cracking in the reddish D_3 interval suggests the existence of a soil formed during subaerial exposure of lake sediments. Relatively wet intervals (W_1 to W_4) precede each of the dry events. Interval W_1 was not recorded in the δ^{18}O data set because the overflow : inflow ratio was too high to permit saturation with $CaCO_3$. Interval W_2 contains two secondary peaks (W_{2a} and W_{2b}).

Although not amplitude-locked, the TIC data parallel oscillations in δ^{18}O throughout the entire time period (Fig. 3). This parallelism indicates that detrital silicates (glacial rock flour) did not obscure the TIC signal, supporting other studies which indicate that the Sierras were essentially deglaciated by 15–14 kyr ago[26,27]. The amplitude of TIC variability increases between 8.8 and 6.7 kyr ago: this suggests that maxima in TIC were caused by carbonate precipitation in extremely shallow lakes. Under these conditions, carbonate precipitation masked the input of detrital silicates whose flux decreased with decreasing inflow of the Owens River. The time of transition to shallow oscillating conditions (8.8 kyr) was coeval with the rapid disappearance of Early Holocene woodlands from the Chihuahuan, Sonoran and Mojave deserts in the southwestern United States ~9 kyr ago[28]. This change to a drier type of vegetation also indicates an increase in aridity of the southwestern United States after 9 kyr.

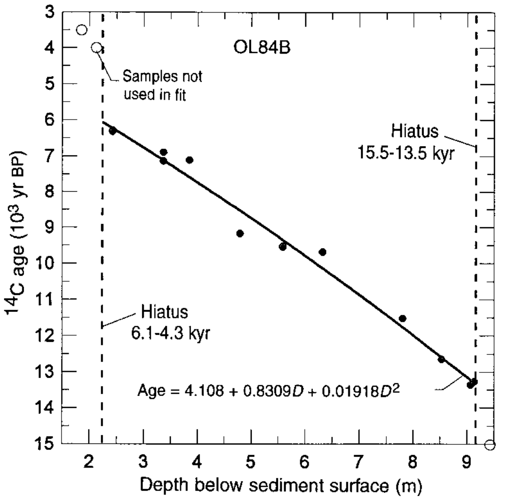

Figure 2 Radiocarbon-age model for core OL84B between depths of 2.25 and 9.20 m. Filled circles indicate ^{14}C analyses on bulk organic carbon used in construction of the age model. Vertical dashed lines indicate location of hiatuses in sedimentation. All age data (kyr) presented in the text of this Letter are in calendar years.

The oscillatory behaviour of TIC was terminated by Desiccation II, which lasted from ~6.7 to 4.5 kyr ago (Fig. 3). Transition to wetter conditions after 4.5 kyr at Owens Lake is consistent with pollen data from meadows in the Sierra Nevada that indicate a transition at ~4.7 kyr to peat and other plant types that require abundant soil moisture[29].

Pollen was extracted from 17 sediment samples from OL84B to determine if rapid changes in the hydrological balance of Owens Lake also were reflected in the vegetation community that surrounded the lake. When the climate was wet and Owens Lake occupied most of its basin, juniper (*Juniperus*) or sagebrush (*Artemisia*) should have composed much of the nearby vegetation. When the climate was dry and Owens Lake was small, the Chenoam (Chenopodiaceae and Amaranthus) group—which includes drought-tolerant salt-resistant desert taxa—should have populated the saline playa abandoned by the retreating lake.

The pollen data indicate that *Juniperus* was abundant between 15.6 and 14.6 kyr ago; thus, it was very wet during the period that followed Desiccation I. After 15.0 kyr, *Juniperus* declined rapidly and then recovered somewhat during subsequent wet phases W_3 and W_4 (Fig. 3). Peaks in the *Artemisia*/Chenoam (A/C) ratio, which indicate relative wet conditions, occurred during minima in the δ^{18}O and TIC records (Fig. 3). Thus all three proxies of climate variability yield a consistent picture of oscillations in the hydrological balance of Owens Lake.

Comparison of the δ^{18}O record of change in the hydrological balance of Owens Lake with the GISP2 record of the Oldest Dryas/Bølling/Older Dryas/Allerød/Younger Dryas and Holocene cold/warm oscillations indicates a remarkable degree of similarity in the number, duration, and timing of climate regimes in both records between 17.0 and 11.2 kyr ago (Fig. 3). There are five North Atlantic cold regimes centred at 16.4, 14.8, 14.0, 13.2 and 12.3 kyr ago and five dry regimes in western North America centred at ≤16.9, 15.1, 14.2, 13.2 and 12.2 kyr ago. Although the errors associated with our age model (±500 yr; Fig. 2) make it impossible to demonstrate absolute synchroneity between the two records, it is clear that both

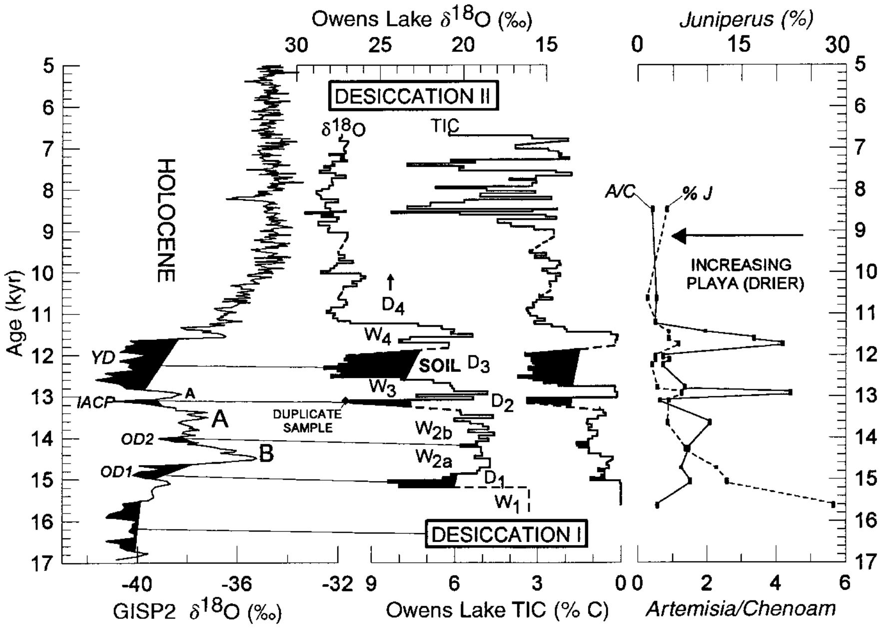

Figure 3 δ^{18}O, total inorganic carbon (TIC) and pollen records for core OL84B. D_1 to D_4 are relatively dry intervals; W_1 to W_4 are relatively wet intervals. The timing of the Oldest Dryas (OD1)/Bølling (B): Older Dryas (OD2)/Allerød (A): Younger Dryas (YD)/Holocene boundaries and the Inter-Allerød Cold Period (IACP) are based on ice-layer counts from GISP2[34]. Cold intervals in the ice-core record and dry periods in the Owens Lake record are indicated by black intervals. Note that the ^{14}C age model for core OL84B is accurate to only a few hundred years. This implies that the δ^{18}O record from OL84B could lead or lag the δ^{18}O record from GISP2; that is, oscillations present in both cores cannot be demonstrated to be synchronous. In the pollen record shown, *A* indicates *Artemisia*; *C*, Chenoam (Chenopodiaceae + Amaranthus) and *J*, *Juniperus*.

records attest to a similar number of large and abrupt climate oscillations during the last glacial termination. We argue that, in general, Atlantic cold events (for example, the Younger Dryas occurred during dry intervals in western North America (for example, D_3); also, warm events in the Atlantic region (for example the Bølling) occurred during wet intervals in western North America (for example, W_{2a}). The last wet phase (W_4) occurred during the last Greenland warm peak.

With the advent of the Holocene, linkage of the climate regimes of the North Atlantic and western North America weakened and perhaps disappeared. Before the Holocene, relatively dry conditions occurred in western North America when the North Atlantic region was relatively cold. During the Early and Middle Holocene this relationship was reversed; that is, western North America was relatively dry and the North Atlantic region was relatively warm.

The duration, timing and similar number of climate oscillations in western North America and the North Atlantic region, indicated by this and other studies[20], suggests a climate-change link during the last glacial termination throughout at least part of the Northern Hemisphere. Errors inherent in our age model do not allow us to completely rule out an oceanic linkage; however, recent climate simulations more strongly support the concept of atmospheric forcing[11,12]. In agreement with these studies, we suggest that oscillations in wetness and temperature in western North America were linked to oscillations in the strength and pattern of the North Atlantic thermohaline circulation through its effect on sea surface temperature and atmospheric water content. Rapid climate oscillations in the North Atlantic regions have been attributed to sudden changes in the rate and location of thermohaline overturn[30–33]. We propose that cooling of the North Atlantic, resulting from a decrease in thermohaline circulation, caused a downstream cooling of the North Pacific[11], which in turn decreased the temperature and moisture content of air passing over the middle latitudes of western North America[12]. ☐

Received 21 October 1996; accepted 29 May 1997.

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Acknowledgements. We thank P. Bartlein and F. Phillips for reviews and suggestions. This work was supported by the US Geological Survey Global Change Program, NSF, and the University of Southern California Faculty Research Innovation Fund. The work was performed in part under the auspices of the US Department of Energy by Lawrence Livermore National Laboratory.

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