Climatic Implications of an 8000-Year Hydrogen Isotope Time Series from Bristlecone Pine Trees

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Tree rings from three dendrochronologically dated bristlecone pines were analyzed for stable hydrogen isotopic composition. These trees give a continuous time series from 8000 years ago to the present that indicates the presence of a postglacial climate optimum 6800 years ago and a continuous cooling since then. The qualitative agreement between this record and records from other sources, such as ice cores, pollen, and treeline fluctuations, indicates that these climate changes were global. This record can serve as a reference for other climate indicators throughout the past 8000 years.

The early and middle Holocene are believed to have been relatively warm, especially in the northern latitudes. A continuous oxygen isotope record showing climatic cooling since 5000 years ago has been reported for an ice core from the Devon Island ice cap in arctic Canada (*1*). However, the Greenland ice cores show much less pronounced trends for the same time period (*2*). Other continental climatic indicators, such as pollen distribution (*3*) and treeline fluctuations (*4, 5*), indicate that summer temperatures were higher between 6000 and 8000 years ago in comparison with the present. Because of the discontinuous nature of most continental climatic indicators, there are few continuous continental records of Holocene climate. Here we present a continuous hydrogen isotopic tree-ring record from specimens from the White Mountains of California of the longest living tree species on earth, the bristlecone pine (*Pinus longaeva*), in order to reconstruct the climatic temperature of the area. The dendrochronology of the bristlecone pine in this area has been well established and dates back to 8000 years before the present (yr B.P.).

The deuterium to hydrogen ratios [expressed as δD (*6*) values] of nonexchangeable hydrogen in the cellulose of tree rings are systematically related to the δD values of the source water used by the tree (*7, 8*). Because the source water is predominantly meteoric water, which is systematically related to climate (mainly to temperature), the δD in trees can be interpreted in climatic terms (*8–11*). The higher the δD in tree rings, the higher the climatic temperature.

The White Mountains are characterized by a cold and dry climate (*12*). In the alpine zone, the average temperature range is 2° to −32°C, and the average annual precipitation is 50 cm. Local variation in precipitation and runoff is strongly influenced by topography. Soil quality is poor, and the growing season is less than 2 months long.

We studied the isotopic compositions of wood sections from three bristlecone pine trees, which we refer to as BrPn1 for the oldest tree, BrPn2 for the tree of intermediate age, and BrPn3 for the youngest tree. They were located at 37°26′N and 118°10′W. These wood sections cover the time periods 6045 to 2220 B.C., 2100 B.C. to 800 A.D., and 180 to 1940 A.D., respectively. The sections were cut into 10-year intervals and ground up, and five sequential samples of equal weight were mixed to give a 50-year interval. We analyzed the cellulose nitrate prepared from the samples, which contains only the nonexchangeable hydrogen of cellulose, using standard procedures (*13*). The analytical uncertainty is ±1.5 per mil (1σ).

The δD values show systematic variations with age (Fig. 1). The three time series are offset from each other. There is an overlap of 700 years between BrPn2 and BrPn3 (Fig. 1A), and for that time period both trees show similar trends in the δD values. The offset between the δD values in the two time series is most likely caused by differences in local environments and growing conditions. There is a gap of 100 years between BrPn1 and BrPn2 (Fig. 1A), and it is unlikely that the isotopic difference between the two trees represents a sudden large climate change. The $\delta^{13}C$ values of these samples show no offset (*15*).

The observed discontinuities in the isotope records may have resulted from nonclimatic conditions. Local topography can play an important role in determining the average δD values of the soil moisture. A hillside may generate more runoff than a flat area, and runoff is mostly generated by summer storms, so the hillside would lose more summer precipitation than the flat area. This results in the source water of plants being more enriched in deuterium in the flat area than on the hillside. At the sampling site, winter precipitation is mainly in the form of snow. Accumulation of snow may vary from one place to another depending on whether it is a ridge or a valley, to windward or leeward. Snow precipitated on the windward slope may be blown back

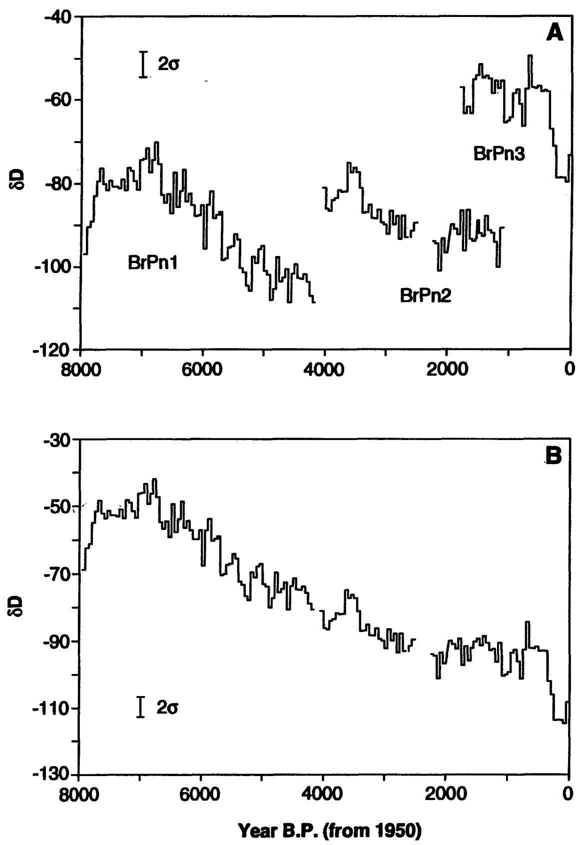

Fig. 1. The δD values of cellulose nitrate plotted versus time in yr B.P. (measured from 1950 A.D.). Cellulose nitrate samples were obtained from wood sections of three bristlecone pine trees. Each data point represents a 50-year average obtained from a mixture of five sequential equally-weighted samples of 10-year intervals. (**A**) The original data set. (**B**) A continuous time series. See the text for the method used to obtain (B) from (A).

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into the air by wind and reprecipitate in a leeward valley. We hence have a case where the leeward valley receives more winter precipitation and, therefore, contains source water that is more depleted of deuterium than that in a nearby windward location. Growing conditions may vary from tree to tree or even within the same tree. Ramesh *et al.* (*14*) showed that asymmetrical growth of a tree can cause a difference of up to 33 per mil in the δD values around a single tree ring. They pointed out, however, that the isotopic signals among different radial directions are coherent. If only the trends or the relative variations are of interest, an arbitrary radial direction of an asymmetrical disc of a tree would contain all potential climatic information carried in the tree. Bristlecone pine trees grow under adverse conditions and their growth patterns are usually irregular. It is likely that the differences in the absolute δD values among the three time sequences have resulted, at least partly, from different growing conditions, which produce effects that are similar to those produced by asymmetrical growth. Nonclimatic conditions affecting growth may also include soil type and conditions and water table depth.

Under the assumption that climate variation in the past was gradual and systematic, a continuous isotopic record on a relative scale was obtained as follows. We subtracted 35 per mil from all the analyses of the samples of BrPn3. The value 35 per mil was selected such that the sum of square of difference between the corresponding δD values of BrPn2 and BrPn3 for the overlapped time period (150 to 850 A.D.) became minimal. After this correction, the average δD values between BrPn2 and BrPn3 were calculated for the corresponding times and used for the overlapped time interval. There is no overlap between BrPn1 and BrPn2. We added 28 per mil to all the analyses of BrPn1, so that the δD value of the youngest sample equals that of the oldest sample of BrPn2. Although the analytical error is only ±1.5 per mil, the uncertainty in the climatic trend is larger because of the necessity of matching different wood segments.

The resulting time series (Fig. 1B) shows a plateau at about 6800 yr B.P. (zero yr B.P. is 1950 A.D.), which correlates with the climatic optimum in the mid-Holocene. There is a general cooling trend from 6800 to about 2000 yr B.P. Between 2000 and 400 yr B.P., the δD curve is fairly flat. A rapid cooling started in 1600 A.D. and the cold climate peaked between 1700 and 1900 A.D. This cold period may correspond to the Little Ice Age. The overall cooling trend shown in Fig. 1B is in good agreement with the records of the treeline fluctuations in the White Mountains studied by LaMarche (*4*). He showed that the upper treelines of bristlecone pine continuously declined to lower elevations from 7400 yr B.P. to about 200 years ago. This adds an important verification to our assumption that the δD records represent a single climate curve.

A long-term decrease in $\delta^{18}O$ (*6*) from 5000 yr B.P. to the present has been reported for an ice core from the Devon Island ice cap in arctic Canada (*1*) (Fig. 2). The general trends of the isotopic curves are similar, although the timing of the climatic optimum shifts from our data by approximately 2000 years. This shift may be due to a phase delay in the onset of cooling at Devon Island or to the large standard dating error for the ice core [20% (1σ) at 5000 yr B.P.]. In consideration of the great differences in the climatic environments and in the natures of the proxy indicators, the agreement in the general trend of these two isotopic records suggests that they both recorded the global climatic changes.

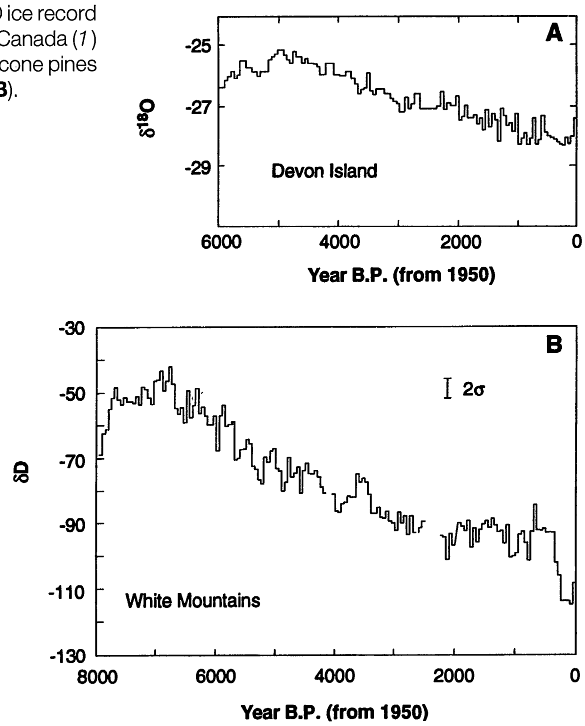

Fig. 2. A comparison between the $\delta^{18}O$ ice record from the Devon Island ice cap in arctic Canada (*1*) (**A**) and the δD record from three bristlecone pines from the White Mountains, California (**B**).

Relations between δD values of tree rings and surface temperature were shown by Yapp and Epstein (*10*) to be 7 to 8 per mil per degree centigrade. The trees they studied were collected from widely distributed geographical locations, from warm places such as Houston, Texas, to cold places such as Alaska. Similar slopes are observed also for the δD and temperature time series of trees from two North American locations, Manitoba and New York state (*15*). With this calibration, the magnitude of cooling from 6800 to 2000 yr B.P. is about 5° to 7°C. However, the δD-*T* (temperature) relation has not been determined for the White Mountains area because of the lack of long-term temperature records. The slope for a correlation between the δD of tree rings, δD_{CN}, and the δD of the source water, δD_S, is a function of relative humidity (*8*). Variations of this slope directly affect the δD_{CN}-T relation. The slope of the δD_{CN}-δD_S line ranges from 0.77 to 0.85 over most commonly encountered humidities, and this is the reason why most of the observed slopes for the δD_{CN}-T line lie between 7 and 8 per mil per degree centigrade. In a dry area, the slope of the δD_{CN}-δD_S line should be approximately 1 (*8*). This slope would correspond to a slope of 5.6 per mil per degree centigrade for the δD_{CN}-T relation, the same slope in the relation between the δD of precipitation and temperature. The White Mountains are relatively dry and might have been drier between 9000 and 6000 yr B.P. (*16*). In conclusion, it is clear that the climate of the White Mountains has been cooling since 6800 yr B.P., and the magnitude of cooling cannot be defined more accurately than within 2° to 3°C.

To a first approximation, the trees recorded the δD values of total precipitation, which reflects the average temperatures of the air masses that bring precipitation to the location. Because each data point is an averaged δD value for 50 years and we discuss only the long-term climatic trend, this approximation may be appropriate. At the present day, precipitation in the White-Inyo Range is from continental polar air, recycled maritime polar air, moist Pacific air, and, less frequently, very moist tropic air from the vicinity of Hawaii (*12*). Once every few years, an invasion of Arctic air from interior Alaska or the Yukon occurs, bringing record cold temperatures. The long-term decrease in the δD values of the precipitations may be a result of changes in proportions of warm versus cold precipitation (for example, summer versus winter precipitation). It is also possible that variations reflect an increase in the frequency of the extremely cold events or a decrease in the frequency of the warm events. This set of observations may, therefore, provide valuable data for the validation of climatic simulations concerning changes in atmospheric circulation patterns.

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6. The δD and $\delta^{18}O$ are defined as

$$\delta D = \left[\frac{(D/H)_{sample}}{(D/H)_{standard}} - 1\right] \times 1000$$

and

$$\delta^{18}O = \left[\frac{(^{18}O/^{16}O)_{sample}}{(^{18}O/^{16}O)_{standard}} - 1\right] \times 1000$$

where standard is standard mean ocean water (SMOW).

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Stream Networks and Long-Term Surface Uplift in the New Madrid Seismic Zone

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Stream networks are sensitive to low rates of surface uplift and can be used to decipher the history of large earthquakes even where faults do not rupture the surface, as in intraplate seismic zones. Statistical analysis of alluvial network data from topographic maps in the New Madrid seismic zone, in the central United States, shows that stream-segment gradients deviate the most from an estimated natural stream profile where surface uplift is greatest. Evidence of cumulative deformation distilled from stream network patterns represents at least several meters of differential surface uplift during Holocene time, which suggests that more than one cycle of surface deformation occurred.

Landform geometry reflects crustal tectonics, and geomorphic features provide a bridge between deformation on the time scales of several earthquake cycles (10^2 to 10^4 years) and recent geodetic surveys (1 to 100 years). Landforms generally analyzed include fault-bounded mountain fronts, offset stream channels, or uplifted coastlines along active plate margins and in deforming mountain belts where faults commonly break the surface. Using landforms to extend prehistoric seismic records in continental interiors where earthquakes are common (*1*) has been unsuccessful because faults rarely rupture the surface in continental interiors and the terrain in such areas typically has little relief. Here, elevation changes in 15 stream networks were analyzed for deformation resulting from the succession of large earthquakes in 1811 and 1812 in New Madrid, Missouri, a continental interior location with historic earthquake activity (*1*).

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The New Madrid seismic zone (NMSZ) spans Quaternary deposits on the borders of Kentucky, Missouri, Arkansas, and Tennessee. More than 1 km of alluvial sediment mantles the continental crust and obscures crustal displacement (*2*). In 1811 and 1812, the three largest earthquakes (magnitude ~8.1 to 8.3) known to have occurred in the continental North American plate emanated near New Madrid (*1*). Although this region is far from a plate boundary, it has had nearly continuous microearthquake activity throughout monitoring in this century (*3*). Shear strain rates are as much as one-third of those along faults forming active plate margins, such as the San Andreas (*4*). Geodetic and geomorphic evidence indicates that tilting is occurring in the midcontinent and might be the cause of downstream changes in channel sinuosity along large rivers (*5, 6*), and paleoseismic investigations of liquefied sands indicate that large earthquakes might have occurred many times during the past several thousand years (*7–9*). Some of the aerially extensive, low-relief features within the NMSZ, such as the Lake County uplift, are related to both active faults and historic seismicity (*8, 10–12*). Other such features either are not or their relation to underlying geologic structures is not yet known (*13–15*).

We studied the Lake County uplift (LCU) region where substantial surface deformation is known to occur and its pattern is clearly established. The LCU is a gourd-shaped, composite Quaternary structure containing anomalous relief (Fig. 1A), active faults, and much seismic activity (*10, 11*). The area of the LCU is about 50 by 23 km and upwarps the Mississippi River valley as much as 10 m. Seismic reflection data indicate that subsurface rocks are upwarped in a manner that parallels the pattern of surface warping (*10, 11*). Seventy-five percent of microearthquakes recorded in the NMSZ between July 1974 and June 1978 occurred within the LCU.

From analysis of warped, active as well as abandoned channels and levees of the Mississippi River, Russ (*10*) divided the LCU into three prominent topographic bulges: Tiptonville Dome, Ridgeley Ridge, and Sikeston Ridge (Fig. 1A) (*10*). We focused on the area encompassed by Tiptonville Dome and Ridgeley Ridge, including their perimeters where Russ's (*10*) estimate of surface uplift tapers to 0. Most of the Tiptonville Dome formed between 200 and 2000 years ago, but additional uplift occurred during the 1811 and 1812 New Madrid earthquakes (*10, 11*). Russ (*10*) estimated that up to 9 to 10 m of surface uplift has occurred, which indicates that rates of surface uplift (4 to 5 mm $year^{-1}$) are as high as those along active plate margins dominated by compression (*16*). Ridgeley Ridge, formed <6000 years ago, is underlain by northeast-trending faults, some of which are possibly high-angle reverse faults, and its surface expression mimics the subsurface structural pattern (*10, 11*).

The northeastern boundary of the LCU