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Dendrochronology and links to streamflow

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SUMMARY

Streamflow variability on timescales of decades to centuries becomes increasingly important as water managers grapple with shortages imposed by increasing demand and limited supply, and possibly exacerbated by climate change. Two applications of dendrochronology to the study of flow variability are illustrated for an existing 1244-yr reconstruction of annual flows of the Colorado River at Lees Ferry, Arizona, USA: (1) identification and climatological interpretation of rare flow events, and (2) assessment of vulnerability of water-supply systems to climatic variability. Analysis centers on a sustained drought of the mid-1100s characterized by persistent low flows on both the Colorado and Sacramento Rivers. Analysis of geopotential height anomalies during modern joint-droughts suggests more than one mode of circulation might accompany joint-drought in the two basins. Monte Carlo simulation is used to demonstrate that a drought as severe as that in the 1100s on the Colorado River might be expected about once in every 4–6 centuries by chance alone given the time-series properties of the modern gaged flows. Application of a river-management model suggests a mid-1100s-style drought, were it to occur today, would drop reservoir levels in Lake Mead to dead-pool within a few decades. Uncertainty presents challenges to accurately quantifying severe sustained droughts from streamflow reconstructions, especially early in the tree-ring record. Corroboration by multiple proxy records is essential. Future improvements are likely to require a combination of methodological advancements and expanded basic data.

1. Introduction

Dendrochronology is linked to streamflow through the common responses of tree-growth and streamflow to variations in net precipitation and runoff (Stockton and Jacoby, 1976). The statistical relationship between time series of tree-ring indices and streamflow has been exploited for multi-century reconstructions of flow for river basins in many parts of the world (e.g., Akkemik et al., 2004; Gou et al., 2007; D'Arrigo et al., 2009; Liu et al., 2010). The methodology and applications of streamflow reconstruction from tree rings have been reviewed by Loaiciga et al. (1993) and Meko and Woodhouse (2011).

Nowhere has the attention to streamflow reconstruction been more focused than in the Colorado River basin, a key source of water supply for some 30 million people in the western United States. Climate change and variability are critical issues in this basin. Mean annual flows of the Colorado River are over-allocated by the 1922 Colorado River Compact governing the distribution of water (MacDonnell et al., 1995); demand is expected to continue increasing (National Research Council, 2007); climate-change projections envision imminent drying over the next century (Seager et al., 2007b); and Lake Mead, a major reservoir, is at its lowest level since it began filling in the 1930s (Barringer, 2010). Projected climate change is expected to result in substantial decreases in runoff and further drops in reservoir levels by the end of the 21st century (e.g., Christensen and Lettenmaier, 2006; McCabe and Wolock, 2007; Barnett and Pierce, 2008; Rajagopalan et al., 2009).

The history of Colorado River water woes is closely linked to the development of the science of dendrohydrology in the United States. Schulman (1945) established the physical rationale for reconstruction in assessing water-supply variability of the Colorado for Los Angeles Power and Light. Stockton and Jacoby (1976) first applied modern multivariate statistical methods to reconstruction in extending the Colorado River flows at Lees Ferry, Arizona to A.D. 1520. Subsequent tree-ring studies of reconstructed flow on the Colorado have aimed at improvement of accuracy, temporal extension, climatological interpretation and water-management applications. Approaches to increasing the accuracy have included varying the makeup of the tree-ring network and exploring new methods of statistical reconstruction modeling (e.g., Michaelsen et al., 1990; Hidalgo et al., 2000; Woodhouse et al., 2006; Gangopadhyah et al., 2009). Climatological interpretation has been directed toward examination of ocean-atmosphere drivers of reconstructed flow variations (e.g., Woodhouse et al.,

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2006, 2010). Water managers have long referred to reconstructions as precautionary evidence of extended droughts on the Colorado, but have only recently explored direct use of reconstructions as input to river-management models (Prairie et al., 2008).

In this paper we analyze the longest existing reconstruction for the Colorado River (Meko et al., 2007) to investigate ways the extended flow record can provide additional information on current and future droughts. We frame the analysis around a particular paleo-drought, a multi-decadal period of recurrent low flows in the mid-1100s (Meko et al., 2007). This low-flow period occurred during a medieval period (A.D. 800–1400) characterized by unusually persistent droughts in western North America and large hydroclimatic anomalies in other parts of the globe (Seager et al., 2007a). The ongoing drought, now entering its second decade (Woodhouse et al., 2010) could represent a return to an amplified low-frequency mode of hydroclimatic variability. Here we explore various aspects of the persistent low-flow period in the mid-1100s on the Colorado River: spatial extent and climatology, water-supply implications, and likelihood of recurrence. We also discuss challenges to interpretation cast by uncertainty. Novel aspects of this paper include a simulation-based approach to a probabilistic context of an exceptional persistent paleo-drought, and application of a long-term river planning model to explore potential impacts on management were such a drought to occur in the future.

2. Data

Tree-ring data points and basins are shown on the map in Fig. 1. Reconstructions analyzed include A.D. 762–2005 annual (water-year) flows of the Colorado River at Lees Ferry Arizona (Meko et al., 2007) and A.D. 869–1977 annual flows of the Sacramento River (Meko et al., 2001); both reconstructed time series were downloaded from the International Tree-Ring Data Bank (ITRDB) (http://www.ncdc.noaa.gov/paleo/treering.html). To illustrate aspects of uncertainly in the Colorado River reconstruction, use is also made of calibration-period (post-1905) segments of sub-period, or time-nested, reconstruction models described in Meko et al. (2007). These time series, not available from the ITRDB, were obtained from the files of the main author.

Historical time series of natural flows (flows adjusted for depletions and reservoir storage) for years 1906–2009 were obtained for the Colorado and Sacramento Rivers in the western United States. Flows summed over the water-year for the Colorado River at Lees Ferry, Arizona, were downloaded from the US Bureau of Reclamation (http://www.usbr.gov/lc/region/g4000/NaturalFlow/). Flows

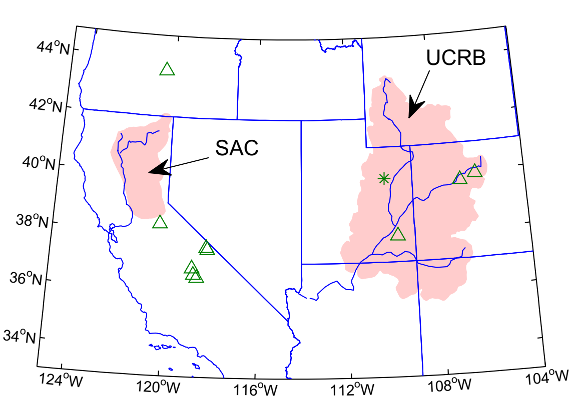

Fig. 1. Map of river basins and tree-ring sites. Sites plotted are Harmon Canyon (∗) and early-1100s tree-ring networks (△) for reconstructions of flow in Upper Colorado River Basin (UCRB) (Meko et al., 2007) and Sacramento River Basin (SAC) (Meko et al., 2001).

for the Sacramento River were downloaded from the California Department of Water Resources (http://cdec.water.ca.gov/cgi-progs/iodir/WSIHIST).[1] All observed and reconstructed natural-flow series used in this paper are included in the Supplemental Materials. Gridded reconstructed summer (JJA) Palmer Drought Severity Index (PDSI) for North America for the past 2000 yr were obtained from Cook et al. (2010).

Unless otherwise noted, flow units on plots are either km^3 or percentage of "normal", defined here as the 1906–2005 observed mean. Some results are also presented parenthetically in million acre-feet (maf) for the benefit of water managers. Normal flow for the Colorado River at Lees Ferry is 18.6km^3, or 15.1 maf; and for the Sacramento River is 22.4km^3, or 18.0 maf.

3. Methods

Low-frequency time series variations were summarized by a 25-yr running mean and by Gaussian filtering. The Gaussian filter is a symmetric, bell-shaped filter with positive weights that sum to 1. We used guidelines in Mitchell et al. (1966) to design a Gaussian filter with approximately the same wavelength of 50% frequency-response (Panofsky and Brier, 1968) as the 25-yr running mean. This is a 33-weight Gaussian filter, which when used for smoothing weights 33 successive years of a time series with highest weight on the central year of the filtered segment; the filter-weights are listed in the Supplemental Material.

Covariation of pairs of time series was summarized by correlation analysis and spectral analysis. Correlation coefficients were tested for significance (Haan, 2002) after adjusting sample-size for effect of autocorrelation (Dawdy and Matalas, 1964). Covariation as a function of frequency was summarized with cross-spectral analysis using the smoothed periodogram as a spectral estimator (Bloomfield, 2000). Steps in spectral estimation were (1) removal of mean, (2) tapering of 5% of each end of the series, (3) padding with zeros to a length the first power of 2 larger than the original series length, (4) raw-periodogram computation by the fast Fourier transform, and (5) smoothing of periodogram with a set of Daniell filters (Bloomfield, 2000) to get a spectrum of desired smoothness and bandwidth. Cross-periodograms and related quantities — squared coherency and phase — were similarly computed using procedures described in detail by Bloomfield (2000) and implemented previously in a tree-ring study by Meko and Woodhouse (2005).

Synthetic time series of observed flow were generated by exact simulation (Percival and Constantine, 2006), using the circulant embedding method of Dietrich and Newsam (1997), to explore how anomalous the most extreme drought of the tree-ring record is given the time-series properties of flow in the gaged record. Exact simulation preserves the spectral properties of the observed series and has the advantage of not requiring an assumption of a parametric generating mechanism (e.g., autoregressive process). Our non-parametric spectral estimator was the raw periodogram, computed as described in the preceding paragraph, except that for the selected circulant-embedding method zero-padding to the next-highest power of 2 larger than four times the original series length was required to supply simulations of the desired length. As normality is an assumption in the exact simulation method of Percival and Constantine (2006), a Lilliefors test (Conover, 1980) was applied to check that time series used to develop simulations are approximately normal. To check sensitivity of probabilities to simulation method we repeated the simulation analysis with synthetic flows generated by a first-order autoregressive (AR(1)), or Markov, process (Haan, 2002). For AR(1) modeling, the model is

[1] This series is referred to online as "Sacramento Valley Runoff", Water-Year Sum

$x_t = \alpha x_{t-1} + \epsilon_t$, where x_t is flow as departure from the mean in year t, α is the autoregressive coefficient, and ϵ_t is the noise term.

Assessment of water-management impact of severe drought was illustrated with the help of the Colorado River Simulation System model (CRSS). CRSS is the official model used by the US Bureau of Reclamation for long-term planning (Fulp, 2003) and represents the Colorado River system using a series of linked objects that interact with each other based on a set of user-specified rules (Zagona et al., 2001). We used the most recent version of CRSS (January 2010), which is current for operating policies and demand schedules. CRSS requires hydrologic input at a monthly timestep for 29 control points located throughout the Colorado River Basin. Accordingly, the Meko et al. (2007) streamflow reconstruction at Lees Ferry was disaggregated both temporally and spatially using the same approach employed by the Bureau of Reclamation (Prairie et al., 2007).

Upper air circulation patterns associated with joint-droughts on the Sacramento and Colorado Rivers were evaluated with composite maps of 500-mb height anomalies. Images of divisional-average PDSI and geopotential height anomalies during joint-droughts were generated with the online Web tool of the NOAA/ESRL Physical Sciences Division, Boulder Colorado (http://www.esrl.noaa.gov/psd/).

4. Return periods of rare flow events

The A.D. 762–2005 reconstruction for the Colorado River at Lees Ferry is a more than tenfold increase over the period of observed natural flows, which begins in 1906 (Fig. 2). One advantage of such extension is an increased chance of sampling uncommon flow features, or rare events, not represented in the short snapshot of observed flows. Extremes are especially sensitive to length of sample. The sample represented by the instrumental period on the Colorado, for example, happens not to include the lowest single-year reconstructed flow, 1685 (Fig. 2). Multi-year droughts consisting of very-low flows clustered over several years, or of slightly-low flows persisting over decades without intervening high flows, are potentially important to water-resources management on the Colorado because such droughts can exhaust the existing reservoir storage. An example of a rare multi-year event of the latter type in the tree-ring record of the Colorado River is the mid-1100s drought highlighted by Meko et al. (2007) as the lowest 25-yr running mean, 15.5 km^3 (12.6 maf or 83.8% of normal), in the 1244-yr reconstruction covering years A.D. 762–2005. In comparison, the lowest 25-yr running mean observed flow in the 104-yr period 1906–2009 was 16.1 km^3 (13.1 maf or 86.9% of normal), which occurred in 1953–1977. The 25-yr window happens to highlight intensity of multi-decadal drought in the Meko et al. (2007) reconstruction, but also has direct relevance to existing management: 1953–1977 is one of two "critical drought periods" designated on the Colorado River by the US Bureau of Reclamation and traditionally used in drought risk-assessment (personal communication, Russell Callejo). [2]

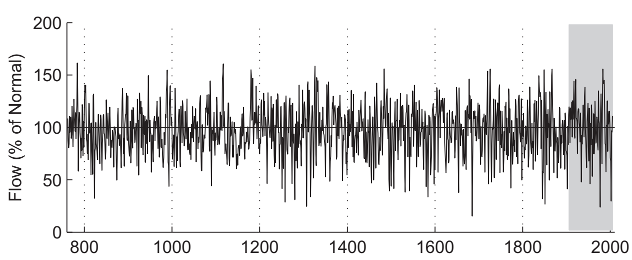

Fig. 2. Time plot of reconstructed annual flows of Colorado River, A.D. 762–2005. Period of available observed flows shaded. Source of data: Meko et al. (2007). "Normal" is 1906–2005 observed mean (see Section 2).

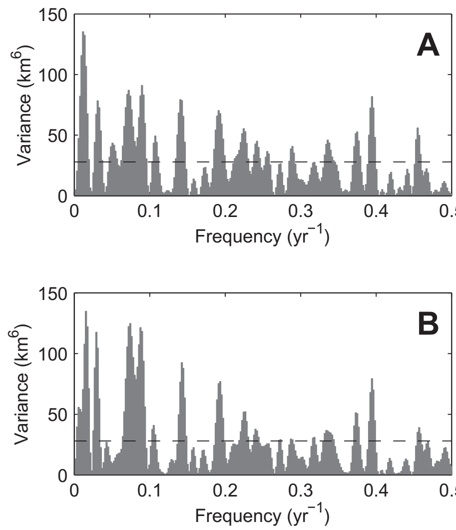

Fig. 3. Periodograms of Colorado River flows used for exact simulations. (A) 1906–1999 observed flows, and (B) 1906–2009 observed flows. Horizontal line at variance of time series (white-noise spectrum).

The mid-1100s drought, can be viewed conceptually as (1) a signal for a changing mechanism of climate/hydrology, or (2) unchanging variability sampled in an expanded time window. Which view is correct is impossible to determine, but simulations can help estimate the chance of those features occurring given the current driving mechanisms. The tree-ring record suggests a probability of $p \approx 1/12$, or $p \approx 0.08$, of a 25-yr mean flow less than 83.8% of normal in any given 104-yr period (approximately 12 such periods in 1244 yr). We defined a "critical drought" as a 25-yr mean lower than 83.8% of normal and used simulations of observed flows to estimate how likely a critical drought would be if many 104-yr sequences of observed flows were available instead of the single snapshot of 1906–2009. The simulation exercise consisted of the following steps:

1. Generate 10,000 simulated series of flows of length 104 yr by exact-simulation (see Section 3)
2. Count the number of simulations N_c with at least one critical drought
3. Estimate the probability of a critical drought in any 104-yr period as $p_c = N_c/10{,}000$

Simulations were repeated on observed flows for 1906–1999 and 1906–2009 to test sensitivity of simulations to the record low-flows of 2000–2009 (Woodhouse et al., 2010). The exercise was repeated by exact simulation and AR(1) simulation to test sensitivity to assumption of generating mechanism. The Lilliefors test indicated both series were approximately normally distributed, and thus suitable for exact simulation. The periodogram, the framework for our implementation of exact simulation, is broadly low-frequency in appearance for both series (Fig. 3). Both series have high variance at $f \approx 0.07$–$0.09\ \mathrm{yr}^{-1}$, or wavelengths 14–11 yr. The effect of the 2000–2009 segment of observed flows is most noticeable as increased variance in that frequency-range and also at $f \approx 0.027$–$0.032\ \mathrm{yr}^{-1}$, or wavelengths 37–31 yr. The underlying shapes of the periodograms (which are unsmoothed

[2] A "critical drought period" is a sequence of dry years, identified from existing flow records, as a worst-case scenario for water management.

spectra), are consistent with autoregressive processes. Both series are positively autocorrelated, and fitted AR(1) models have estimated coefficients, $\alpha = 0.24$ for 1906–1999 and $\alpha = 0.27$ for 1906–2009, that differ from zero by more than two standard errors.

The exact simulations give higher probability of critical drought than suggested by the one occurrence in the 12-century reconstruction, and suggest one such drought might be expected about once every six centuries ($p \approx 0.17$) from 1906 to 1999 flows, and once every four centuries ($p \approx 0.28$) from 1906 to 2009 flows (Fig. 4). The higher probability of critical drought for the longer observed record probably reflects the greater low-frequency variance evident in the periodogram for the 1906–2009 series (Fig. 3) and possibly also the slightly lower mean flow of the 1906–2009 segment (18.50 km^3 versus 18.63 km^3). AR(1) simulation gives lower probabilities than exact simulation. This difference suggests simulation method can make an important difference in estimated probability of critical drought. Exact simulation has the advantage of flexibility in modeling the complicated spectral structure of observed flows (Fig. 3), and for that reason we favor the exact-simulation results over those from AR(1) simulation for these particular time series. For the 1906–2009 segment, however, even the AR(1) simulations yield a higher probability of critical drought than suggested by the long-term reconstruction.

Simulation results suggest the extreme reconstructed 25-yr mean flow of the mid-1100s has a reasonably high chance of recurrence without invoking changes in the modern statistical framework of observed flows. Failure to observe such a drought in the 1906–2009 record is likely a phenomenon of a short sample. A caveat is that the probabilities associated with the simulation exercise just described depend the specific definition of critical drought. A low 25-yr running mean is just one characteristic of the mid-1100s drought described by Meko et al. (2007). Simulation exercises defining drought differently (e.g., number of consecutive years without any "high" flows) could lead to different assessments of "uniqueness" of the 1100s paleo-drought.

Future tree-ring collections of ancient wood in the Colorado basin will likely lead to improved accuracy of flow reconstruction and possibly to re-assessment of drought severity in the mid-1100s. The model generating the A.D. 762–1182 portion of the reconstruction plotted in Fig. 2 has $R^2 = 0.60$. A reconstruction with a higher R^2 would have higher variance, and so a tendency for greater extremes of running means, and this might be expected to yield a lower estimated smoothed flow in the mid-1100s. This result is not guaranteed, however, as lower error variance does not imply that *any particular* reconstructed time series feature will be amplified as a departure from the mean.

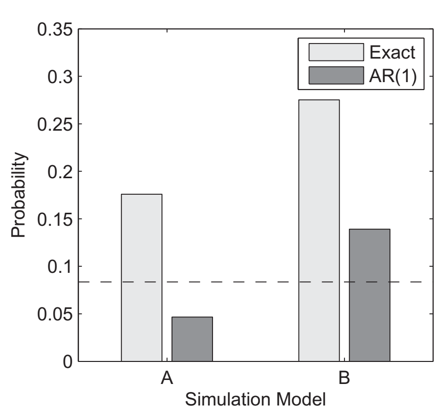

Fig. 4. Simulation-based probabilities of an 1100s low-flow in any 104-yr period. Simulations by exact simulation and first-order autoregressive (AR(1)) modeling of (A) 1906–1999 observed flows, and (B) 1906–2009 observed flows. Dashed line marks empirical probability from 1244-yr tree-ring reconstruction. See Section 4 for definition of 1100s low-flow.

5. Spatial aspects and mechanisms of severe drought

Clues to the climatology of the mid-1100s Colorado River drought are provided by the rich archive of millennial-length tree-ring data for the western United States. In this section we exploit two particularly useful products of this archive: an A.D. 869–1977 reconstruction of the annual flows of the Sacramento River (Meko et al., 2001), and gridded reconstructions of PDSI in the North America Drought Atlas (Cook et al., 2007, 2010). A similar approach to studying mechanisms of past drought in the region on a shorter time scale has previously been taken by Meko and Woodhouse (2005).

The flow reconstructons for the Sacramento and Colorado rivers cover a common period of A.D. 869–1977 and were made possible by the profusion of long drought-sensitive tree-ring chronologies in the Sierra Nevada of California and the Rockies of Colorado. The annual flows of the two rivers are significantly positively correlated, though correlations are small (Table 1). Inter-basin correlation is similar for the observed and reconstructed flows in the 1906–1977 period in common for both types of data. For the longer reconstruction common period (A.D. 869–1977) correlation drops slightly but is even more statistically significant due to the increased sample-size.

Flow anomalies at the decadal scale are large enough to be practically significant in these basins. For example, the range in 25-yr running means of observed flows expressed as a percentage of the 1906–2005 mean is 87–117% on the Colorado and 83–110% on the Sacramento. Smoothed time plots of reconstructed flows show occasionally large in-phase and out-of-phase behavior (Fig. 5). A similar picture of covariation is given by the running mean and Gaussian smoothing. Both smoothed series flag the mid-1100s periods for unusually strong spatial coherence of low flows.

Joint-drought in these two large basins suggests an associated large-scale anomaly in atmospheric circulation and moisture delivery, presumably in the cool season, as the hydrology of both basins is strongly snowmelt-driven. Because multi-decadal joint-droughts are not represented in the 20th and 21st centuries, climatological data for a diagnostic study is lacking. Nevertheless, the mechanism of persistent joint-drought in these critical water-supply basins is an important topic for climatology and hydrology, and the phenomenon of joint-drought at shorter time scales can be investigated with upper-air geopotential height data extending back over more than 50 yr.

While the upper Colorado and the Sacramento River basins share 1977 as the most severe single-year drought in the instrumental period, the most severe multi-year drought occurs at different times in the two basins: the 2000s in the Colorado, and the 1930s in the Sacramento. The late 1980s to early 1990s, however, was a significant period of persistent low flow in both basins. During the period 1988–1992 average annual flows were 72% of

Table 1
Pearson correlation between annual flows of Sacramento and Colorado Rivers.[a]

	Period	r[b]	N[c]	N_e[d]	p-value[e]
Obs	1906–2009	0.44	104	99	5.0E−6
	1906–1977	0.37	72	70	1.5E−3
Rec	1906–1977	0.32	72	70	7.6E−3
	869–1977	0.22	1109	1048	1.0E−12

[a] Natural flows, as described in Section 2.
[b] Correlation over indicated period.
[c] Length of period (yr).
[d] Effective length of period (adjusted for lag-1 autocorrelation).
[e] Probability of Type I error in two-tailed test of null hypothesis of zero population correlation.

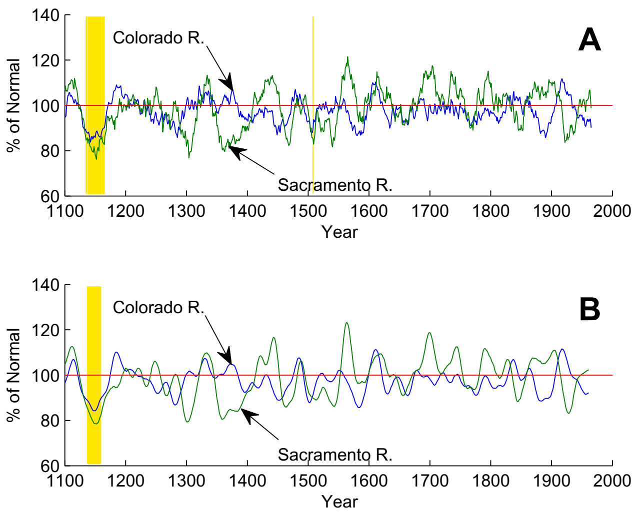

Fig. 5. Smoothed reconstructions of annual flows of Colorado River and Sacramento River. (A) Smoothing by 25-yr running mean. (B) Smoothing by 33-weight Gaussian filter. Periods two series simultaneously in lowest decile shaded. "Normal" is 1906–2005 observed mean (see Section 2). Data sources: Meko et al. (2007) and Meko et al. (2001).

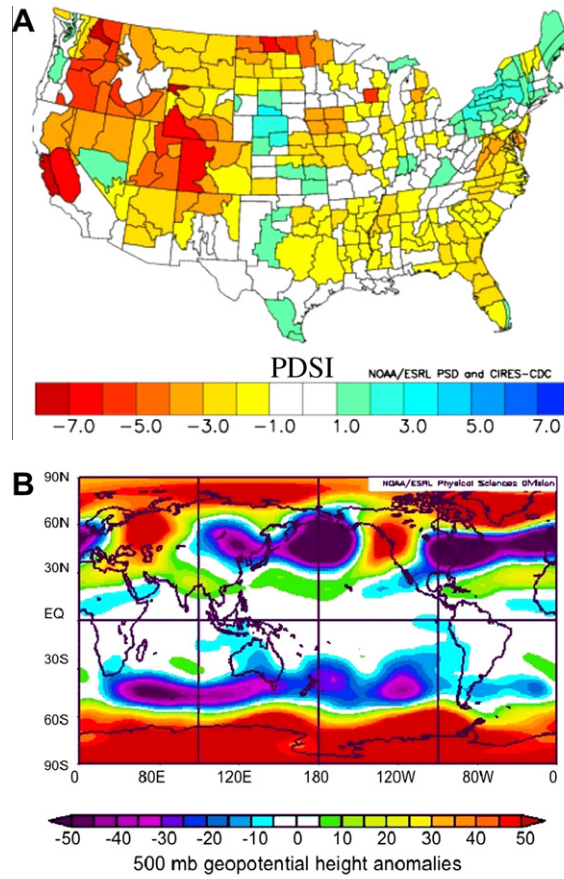

Fig. 6. Drought pattern and circulation anomalies in 1977. (A) Climate Division summer (JJA) Palmer Drought Severity Index, (B) November–March 500 mb geopotential height (m) anomaly from 1968 to 1996 average. Images provided by the NOAA/ESRL Physical Sciences Division, Boulder, Colorado from their Web site at http://www.esrl.noaa.gov/psd/.

normal in the upper Colorado River basin and 56% of normal in the Sacramento River basin.

The North America Drought Atlas (Cook et al., 2007, 2010) can be used to compare spatial patterns of drought in recent decades with patterns reconstructed from tree rings for the mid-1100s. Despite a mismatch of seasons, summer-average (JJA) PDSI patterns are relevant to streamflow in the snowmelt-driven Colorado and Sacramento basins because in the southwestern USA summer PDSI is strongly correlated with cool-season moisture anomalies (Cook et al., 2007; St. George et al., 2010). The drought of water-year 1977 by this summer PDSI metric was widespread across the western USA (Fig. 6a). Atmospheric circulation over the cool season was characterized by a meridional flow pattern with a strong center of high pressure over Pacific Northwest coast of the USA, extending over much of western North America (Fig. 6b). This area of high pressure effectively blocked the jet steam from most of the USA except for far southern portions.

The drought of the 1988–1992 displays a pattern of dryness that extends from California northeast to the northern Great Plains (Fig. 7a). Colorado is on the fringe of this drought, but the headwaters of the Colorado River clearly experienced its influence. Circulation patterns for this period of time are marked by a band of high pressure across eastern Asia, the North Pacific Ocean, and North America (Fig. 7b). Low pressure is restricted to high northern latitudes. A broad ridge is located over the Pacific coast, and the path of the jet stream is over northern North America. The onset of this drought coincided with a cold ENSO event, with below average sea surface temperatures in the equatorial Pacific and the jet stream position north of its typical path (Trenberth et al., 1988). These cold equatorial Pacific conditions did not persist, but the high pressure over western North American continued throughout the years of the drought. This period was also one of widespread drought over much of Europe, with unprecedented impacts on aquifers in lowland areas of the United Kingdom (March and Monkhouse, 1993) perhaps due to strong positive phase of the Northern Atlantic Oscillation.

An exploratory analysis of pattern correlations of reconstructed-PDSI maps for the mid-1100s with observed-PDSI maps of the 20th century did not identify the joint-drought years of 1977 or 1988–1992 as especially strong analogs for years 1138–1160 (joint-drought core in Fig. 5). In general, drought over this interval is widespread through the southwestern and southern USA, although a few individual years (e.g., 1147 and 1154) show the diagonal pattern of drought from southern California to the north-central USA (Fig. 8). The Atlas may however be poorly suited for this type of diagnostic analysis because of limitations in the tree-ring site coverage in the 1100s (in particular, data are lacking for the northern Great Plains, a center of drought in 1988–1992) and the blurring of seasonal climate anomalies over multiple seasons in summer-average PDSI. In summary, the two severe short-period joint-droughts (Colorado–Sacramento), which share somewhat similar patterns of drought, and for which upper air height anomalies are available, are characterized by strongly differing patterns of 500-mb height anomaly. That the circulation patterns are dissimilar points to the complexity of unraveling causes of drought. The general drought pattern for 1977 and 1988–1992 is not common during the 1100s drought, which may be largely due to sparseness of data in key regions. Whether controls on drought in the mid-1100s were analogous to controls on more recent droughts is difficult to say, but may become more evident with increased tree-ring coverage over North America through the medieval period.

6. Water-supply vulnerability

Large river basins in semi-arid regions are often so heavily developed that the impact of climate variation on water supply can be estimated only with the help of river-management models that incorporate the various components of water transfer, storage

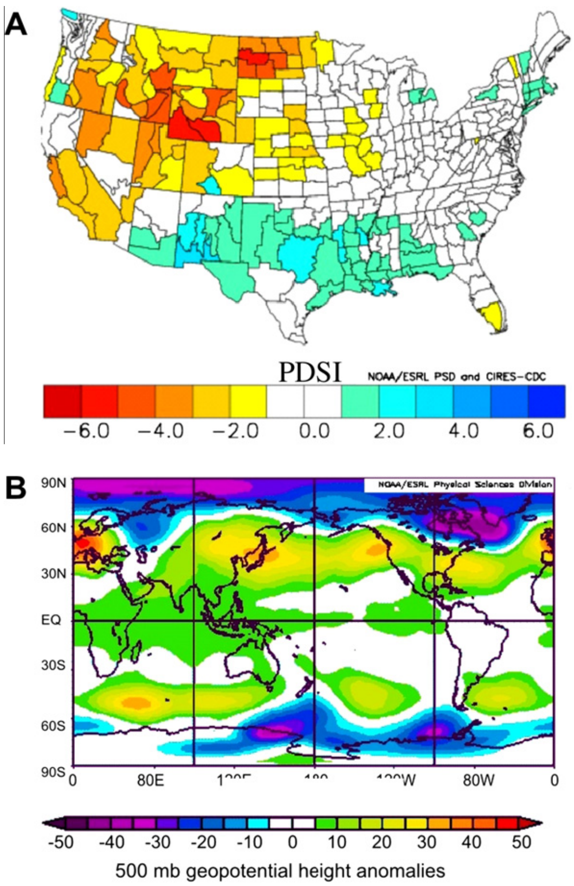

Fig. 7. Composite drought pattern and circulation anomalies in 1988–1992. Remainder of caption as in Fig. 6.

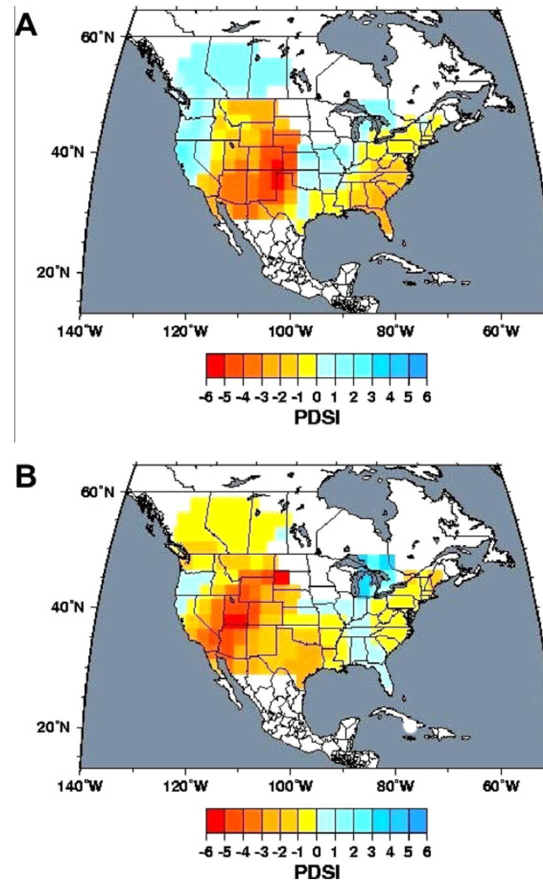

Fig. 8. Tree-ring-reconstructed PDSI in two years of joint Colorado–Sacramento drought of mid-1100s. (A) 1147, (B) 1154. Gridded summer-average (JJA) PDSI from Cook et al., 2010.

and demand. On the Colorado River, the official model for long-term planning is the Colorado River Simulation System model, or CRSS (Fulp, 2003). This model can be used to simulate how system storage responds to various hydrologic inputs.

One approach to using paleo-data in water-management models is to run scenarios from the flow reconstruction as input sequences. In assessing drought sensitivity, it is desirable to choose a period of anomalously low flow. The severe drought of the 1100s has already been mentioned as a rare multi-decadal feature in hydrologic drought in the western United States. A color-mapped plot of anomalies in running means of reconstructed flow highlights this period, and indicates the unique feature of that drought is longevity rather than short-period intensity (Fig. 9).

We selected the 25-yr period with the lowest mean streamflow as input to CRSS and targeted Lake Mead elevation over the management horizon of 2010–2034 as an indicator of system response. To put this extreme drought into perspective, we also selected the 49 next-lowest 25-yr periods as input into CRSS. Mean streamflow for this subset of 50 paleo-data sequences ranged from 15.5 km^3 (12.6 maf) to 16.3 km^3 (13.2 maf), or from 84% to 88% of normal flow. Low-flow periods occurred in seven distinct groups of overlapping sequences which were distributed throughout the paleo-record (see Fig. 9 where orange streaks intersect the dashed line). The maximum level of overlap within groups containing at least two sequences was 96% (when sequences where offset by a single year); whereas, the minimum level ranged from 8% to 92%.

Lake Mead response to the extreme drought sequence of the 1100s is an initial rapid decline in elevation; two of three shortage trigger elevations are breached within the first 5 yr (Fig. 10). By the early 2030s, Lake Mead elevation is approximately at dead-pool elevation (273 m, 895 ft) and remains there until the end of the analysis period. The ensemble of traces produced by the other sequences show that the hydrologic input from the extreme drought sequence does not consistently result in the lowest Lake Mead elevations (Fig. 10), indicating that more intense short-term droughts are embedded within the other low-flow sequences. Relative to these sequences, however, the mid-1100s drought stands out as having no periods of substantive recovery; Lake Mead elevations are either declining or stable.

7. Uncertainty

Uncertainty is an obstacle to the acceptance of streamflow reconstructions by water managers and to the application of reconstructions in water-resources research. The Colorado River reconstruction of Meko et al. (2007) happens to be a relatively strong regression-based reconstruction: reconstructed flows closely track observed fows, and accuracy as measured by regression R^2 ranges from 77% for the segment beginning in A.D. 1365 to 60% for the segment beginning in A.D. 762 (Fig. 11). The degradation of signal back in time is characteristic of reconstructions generated by time-nested models and is a natural consequence of sparse sample coverage of the watershed in the early part of the tree-ring record. The limitation on inferences of flows in the mid-1100s can be appreciated from the tree-ring networks shown in Fig. 1. The Sacramento is problematic because none of the mid-1100s tree-ring sites are within the basin. The tree-ring sites for the Colorado

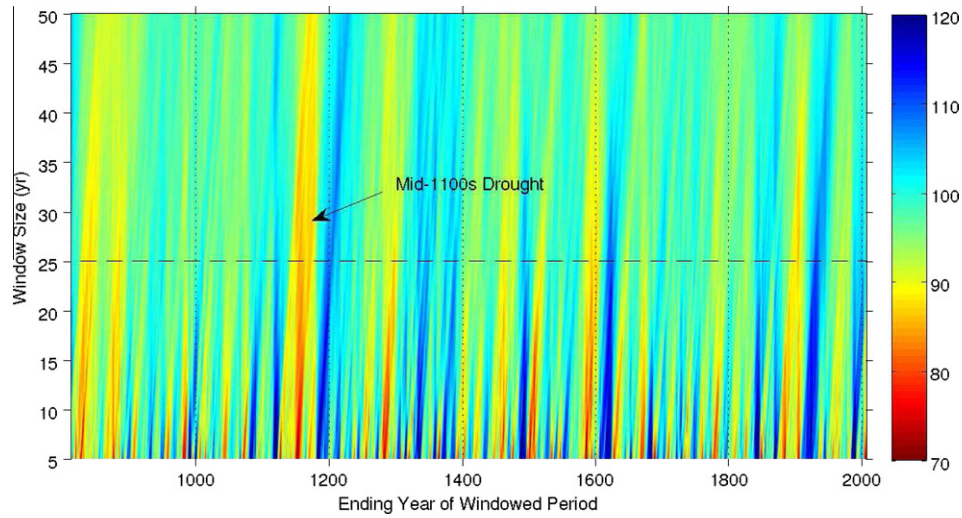

Fig. 9. Color-mapped running means of reconstructed Colorado River flow. Running means of length 5–50 yr mapped on ending year of period. Periods of low 25-yr runnings, say, correspond to yellow-red patches along the dotted line. Color-mapped quantity is percentage of 1906–2005 observed mean (Section 2). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

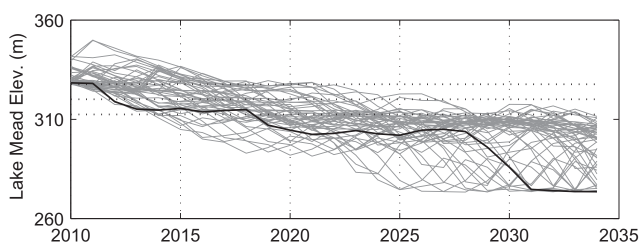

Fig. 10. Ensemble of traces showing Lake Mead elevations for a set of hydrologic input sequences from the paleo-record. Bold line is the 25-yr period with the lowest mean streamflow; other traces show Lake Mead response to 49 25-yr periods with the next-lowest mean streamflow. Shortage trigger elevations are shown by dotted lines and signal delivery reductions to the Lower Basin states in amounts defined by the most recent Colorado River management guidelines (US Department of the Interior, 2007).

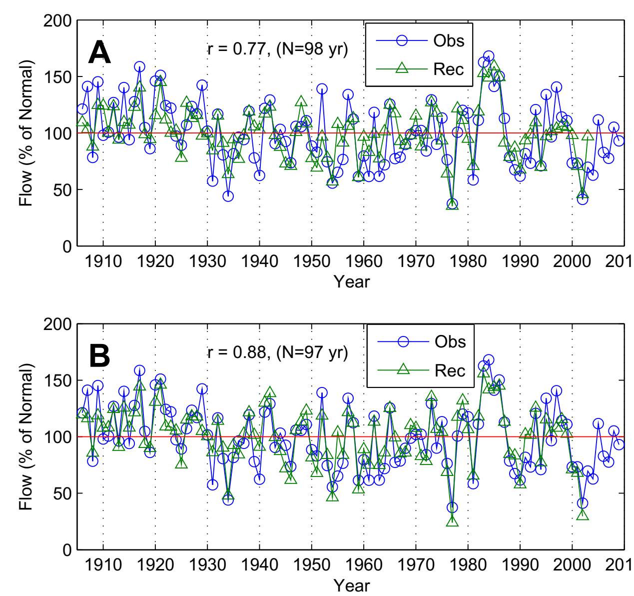

Fig. 11. Time plots of observed and reconstructed Colorado River flows for two sub-period reconstruction models. (A) M762, calibrated on 1906–2003 and used for flows A.D. 762–1182. (B) M1365, calibrated on 1906–2002 and used for flows A.D. 1365–2002. Units are percentage of observed mean (see Section 2). Correlation for period of calibration annotated.

reconstruction are somewhat more favorably situated along the axis of the main stem of the river, but the Green and San Juan sub-basins are not represented in the 1100s network.[3]

When low-frequency features, such as the 1100s drought, are of primary interest, it is important to check that the accuracy implied by regression R^2 applies to the low frequencies. A coherency plot from cross-spectral analysis of observed and reconstructed flows shows that the signal at low frequencies is relatively strong for Colorado River model M762 (the applicable model for the mid-1100s part of the reconstruction), and that frequencies with high coherency generally are those with high variance in the individual series (Fig. 12). The phase diagram in Fig. 12 further supports that variations at those frequencies corresponding to wavelengths longer than about a decade are in-phase. This result lends some support for the ability of the tree-ring record to reflect persistent droughts like that of the A.D. 1100s.

It is important to emphasize that R^2 and other regression statistics do not summarize all aspects of uncertainty in a tree-ring reconstruction of streamflow. It has long been recognized that the detrending of measured ring-width series in standardizing tree-ring data puts a limits on the maximum wavelength of climate variation that can be identified (Cook et al., 1995). Climate fluctuations at wavelengths longer than the longest lifespan of individual trees in a chronology will consequently be missing from tree-ring data standardized by conventional methods. Improvement in identification of low-frequency climate signals is an active research topic in dendroclimatology (Esper et al., 2003; Melvin and Briffa, 2008; Briffa and Melvin, 2011).

Regression statistics also fail to account for possible differences in the sample-depth and makeup of tree-ring chonologies in the calibration period of the regression model and the distant past. This point is illustrated for Harmon Canyon, a *Pseudotsuga menziesii* chronology in northeastern Utah (Fig. 1). None of the sampled trees in this chronology cover both the 20th century and the period prior to A.D. 1200, and sample-depth drops off consideraby in the early part of the tree-ring record (bottom, Fig. 13). It is necessary in

[3] The maps in Fig. 1 show only those tree-ring sites available for the mid-1100s parts of the reconstructions. Later parts of reconstructions have increased site density.

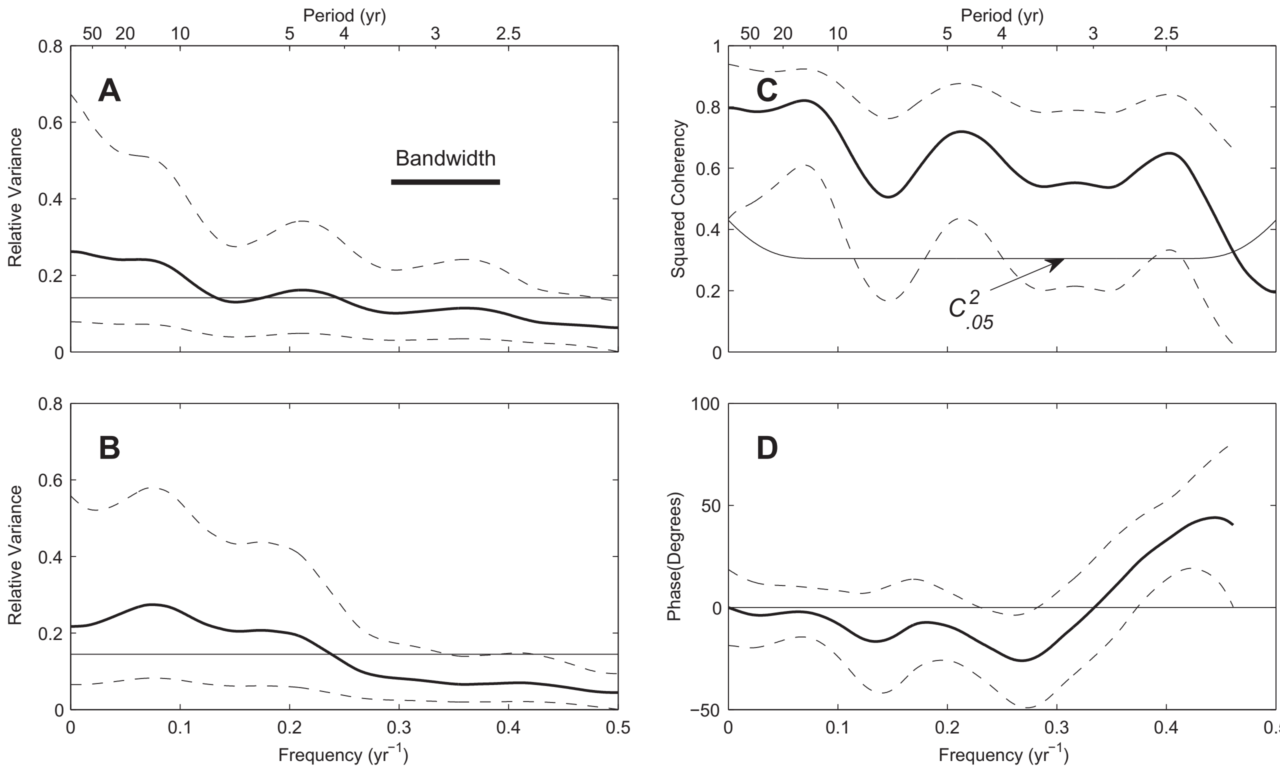

Fig. 12. Cross-spectrum of observed with reconstructed Colorado River flows, 1906–2005. (A) Normalized spectrum of observed flows. (B) Normalized spectrum of reconstructed flows. (C) Squared coherency of observed with reconstructed flows. (D) Phase of observed with reconstructed flows. Horizontal lines in (A) and (B) are theoretic white-noise spectra. Solid line labeled $C^2_{.05}$ in (C) is simplified threshold for 95% significance of squared coherency (Bloomfield, 2000). Dashed lines in all plots are 95% confidence intervals.

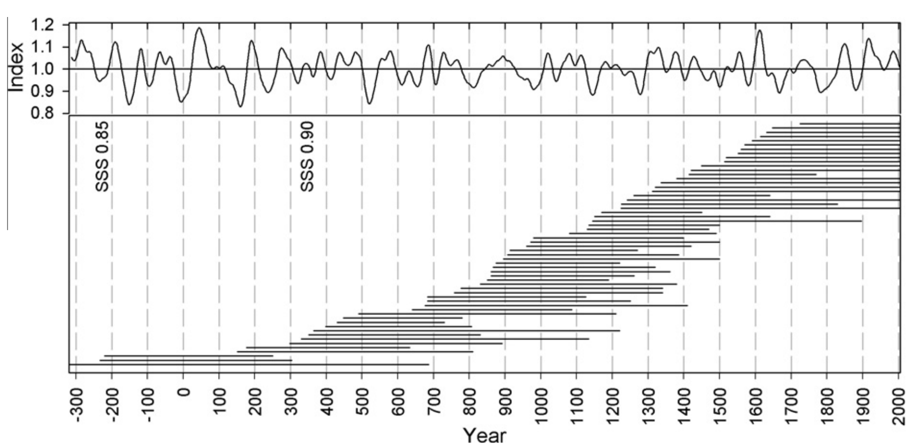

Fig. 13. Smoothed time series and sample-depth chart for Harmon Canyon, Utah, tree-ring chronology. (Top) Standard tree-ring index smoothed with spline to emphasize multi-decadal departures. (Bottom) Time coverage by individual tree radii. SSS, or sub-sample signal strength, measures ability of chronology with given sample-size to capture signal of chronology with full sample-size. Figure from Knight et al., 2010.

applying such a chronology to streamflow reconstruction to assume that the quality of streamflow signal is similar in the living trees and the remnant-wood samples, and reasonable to assume that uncertainty will be amplified in those parts of the record represented by few trees.

An important part of the streamflow-reconstruction process is independent validation of important features of flow reconstructions. This process usually includes reference to other paleoclimatic records and reconstructions in the region (e.g.,Meko et al., 2007). The Harmon Canyon tree-ring chronology plotted in Fig. 13 (top) corroborates the mid-1100s Colorado River drought in that tree-growth at Harmon Canyon then was lower than at any time from the early A. D. 500s to present. This is independent validation because the Harmon Canyon chronology prior to A.D. 1183 was not used in the streamflow reconstruction itself (Meko et al., 2007). An early A.D. 500s low in growth at Harmon Canyon is intriguing (Fig. 13 , top), but without data from other parts of the Colorado River basin cannot yet be interpreted as evidence of basin-wide drought.

8. Conclusion

Robust and accurate estimates of the magnitude and duration of severe hydrologic droughts is an important contribution of dendrochronology to hydrology and water-resources management.

Analysis of a persistent reconstructed drought in the mid-1100s on the Colorado River illustrates both the type of information tree-ring analysis can provide, and the challenges to interpretation. The 1100s drought is a rare event in that no similarly low 25-yr running mean of flow exists in the observed record of natural flows. Time series of simulated flow suggest, however, that given the persistence properties of observed flows we might expect as low a 25-yr mean flow on average about once every 4–6 centuries. Our simulation results are consistent with the mid-1100s drought as a footprint of today's flow variability viewed in an expanded time window. In that sense the magnitude of the mid-1100s tree-ring drought cannot be regarded as remarkable. Impact on multiple large basins in the western United States is an aspect of the 1100s drought that deserves increased attention in hydrologic and climatological studies. On the Colorado alone, the drought is shown using the CRSS model to have devastating potential impact in lowering lake levels.

With the mid-1100s drought we are addressing primarily a low-frequency climatic signal in the tree-ring record. The mid-1100s was in fact not unusual for severity of low flows in individual years, but for the persistence of moderately-dry years and absence of wet years for several decades (Meko et al., 2007). The methods used here can be generalized to the study of higher-frequency features. Higher-frequency fluctuations in runoff can also be important to water resources in the Colorado River basin, and depending on the hydroclimatology, storage, and management practices, may be of primary importance in other basins.

Reducing uncertainty of streamflow reconstruction at all frequency ranges remains a great challenge, especially in the early part of the tree-ring record. The Colorado River reconstruction has a demonstrated signal at multi-decadal wavelengths, but multi-century wavelengths are problematic. Reduced uncertainty for periods more than 800–1000 yr ago is likely to require a combination of improved statistical modeling procedures and — for large basins such as the Colorado — better data coverage by tree rings.

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Appendix A. Supplementary material

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.jhydrol.2010.11.041.

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