Chapter 2:

Significant change-points of subperiod levels in tree-ring chronologies as indications of climate changes

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Draft Abstract (150-500 words):

This chapter introduces three algorithms. First, the scanning $t$-test is used for detecting significant changes in subperiod levels through tree-ring chronologies on multiple time-scale. Second, the scanning $F$-test is used for detecting significant changes in subperiod variances on corresponding scales. Third, an algorithm was developed for comparisons to the palaeoclimate estimates from cored sediments with unequal intervals.

Applications of these algorithms is exemplified for two multi-millenial tree-ring based temperature and precipitation proxies, from Finnish Lapland and western USA, respectively, covering the past 7.6 and 8.0 thousand years. The scanning t-test results obtained from the temperature proxy of Finnish Lapland tree-ring chronology are further compared with palaeoclimatic evidence from ice-core and glacier data from Greenland and Alps, respectively. Moreover, by using coherency analysis, the t-test results obtained from the precipitation reconstruction of the Nevada Climate Division 3 region in the western USA are compared with that in sedimentary data ($\delta^{18}O$, TIC) of Pyramid Lake in Nevada.

It is shown that the scanning $t$-test and scanning $F$-test may provide useful evidence of past and present climatic changes to detect the statistically significant changes in tree-ring based palaeoclimatic records. Moreover, the multi-proxy comparisons make it possible to broaden the view of regional climate variability over larger areas to enhance our understanding on local to near-global climate changes.

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§1 INTRODUCTION

The research topic of changepoint is now applied to many types of digital series in various natural, and even in societal sciences [Wikipedia 2014]. Detections of change-points of statistical parameters in time series (temporal data) could be recall back to a prominent pioneering study by Page [1955] in the statistical literature. Then, Hinkley [1969, 1971] applied regression method to infer a change in two-phase intersection. Pettitt [1980] present a cumulative sum method to detect the level changepoint. Other good works, for example, Kander and Zacks [1966], Brown et al. [1975], Hawkins [1977], and Chen and Gupta [2000], etc.


However, all of the studies mentioned above focused on only searching one change point of subseries means or trends per calculation in a time series. Another type of analyses were introduced in Foufoula-Georgiou & Kumar [1994], they adopt the wavelet transform to detect multi-points of sharp transition, or micro-fronts in the atmospheric turbulence, on various time scales in a long signal series and the wavelet coherency analysis. Unfortunately, they did not give any threshold at a statistical confidence level to check if it is a significant change point at that time [Foufoula-Georgiou & Kumar,1994].

Jiang, et al. [2001, 2002, 2003, 2007, 2008] developed algorithms of the scanning statistic tests for detecting significant change points of subpeoriod means and variances and applications to the climate and hydrologic data. Then Jiang [2009] presented the scanning detection formula for the four parameters, i.e. subperiod means (the first moment), the scanning F-test for detecting changes of subsseries variance (the second moment), trends, and correlation coefficients of a pair series, as well as adding the correction of dependence in the time series to be tested. The scanning test algorithms are based on grafting the wavelet technique onto the traditional Student t-test and F-test correspondently. In this chapter, only the multiscale change points of subpeoriod means and variances are detected and analyzed.

§2 METHODOLOGY

2.1 TWO STATISTIC PARAMETERS IN CLIMATE CHANGE

The IPCC [2001] proposed two statistic parameters in climate change, the mean or average (i.e. the first moments), and the standard deviation or the variance (i.e. the second moments). Suppose a meteorological element, e.g. air temperature, in the Normal distribution, when the subperiod averages change towards higher, the probability of extreme high events (climate disasters) increases, while probability of extreme low events (opposite climate disasters) decreases, otherwise versa (Figure 2.1). On the other hand, when the subsseries variances change towards larger, the probabilities of both side of extreme events (climate disasters)
increases, and otherwise versa (Figure 2.2). In this chapter, we attempt to introduce and apply algorithms for detecting both aspects of significant climate changes. These are also applicable for other digital time series.

2.2 SCANNING T-TEST

Here, the statistic $t(n,j)$ of scanning $t$-test is defined for detecting significant changes of subseries means, by using the traditional Student $t$ parameter for the difference of subseries means between two subperiods in the same one long time series, but with self-variable of the subseries sample number $n$, and the time point $j$. The algorithm formula are as follows (Jiang, et al. 2002, 2003, 2007):

$$ t(n,j) = (\bar{x}_{j2} - \bar{x}_{j1}) \cdot n^{1/2} \cdot (s_{j2}^2 + s_{j1}^2)^{-1/2}, $$

where

$$ \bar{x}_{j1} = \frac{1}{n} \sum_{i=j-n}^{j-1} x(i), \quad \bar{x}_{j2} = \frac{1}{n} \sum_{i=j}^{j+n-1} x(i); $$
| $s^2_{j1} = \frac{1}{n-1} \sum_{i=j-n}^{j-1} (x(i) - \bar{x}_j)^2$; $s^2_{j2} = \frac{1}{n-1} \sum_{i=j}^{j+n-1} (x(i) - \bar{x}_j)^2$. |

Where the subsample size $n$ may vary as $n=2, 3, ..., N/2$, or may be selected at suitable intervals. The $j=n+1, n+2, ..., N-n+1$ is the reference time-point, at which a significant change-point is to be tested on the time scale $n$ in the time-series with total records $N$.

As the series examined in this chapter are somewhat auto-correlated, for example, in the precipitation reconstruction (see §4) the lag-1 auto-correlation coefficients vary between -0.29 and +0.26 in pooled subsamples of length $2n$. The Table-Look-Up Test [Von Storch & Zwiers, 1999] is adopted to correct the significance criterion of the statistic $t(n, j)$ according to the lag-1 auto-coefficients of the pooled subsample in $2n$, and the subsample size $n$. Criterion $t_{0.05}$ after correcting the dependence is usually selected to determine significant changes on time-scales longer than 30 years. For shorter sub-sample sizes, the critical values are usually overly restrictive.

Since the significance level varies with $n$ and $j$, to make values comparable, the test statistic is normalized as:

$$t_i(n, j) = t(n, j)/t_{0.05}$$

The results of the scanning $t$-test are the values of $t(n,j)$ in 3-dimension of the time-scale ($n$), the time ($j$) and the $t_i$ values. The detection of significant points will be completed by excerpting the comparative local maximum or minimum centers/points for $t(n,j)>1.0$ indicating a significant increase change, or for $t(n,j)<-1.0$ denoting a significant decrease change, respectively.

Finally, a coherency index of significant changes between two series $u$ and $v$ is defined as the statistic $t_{nc}(n, j)$:

$$t_{nc}(n, j) = \text{sign}[t_{nu}(n, j) \cdot t_{nv}(n, j)] \{t_{nu}(n, j) \cdot t_{nv}(n, j)\}^{1/2}$$

Usually, a local maximum center of $t_{nc}(n, j)>1.0$ with both $|t_{nu}(n, j)|$ and $|t_{nv}(n, j)|>1.0$ represents a pair significant changes in a same-direction or in-phase, while a local minimum center of $t_{nc}(n, j)<-1.0$ denotes a pair significant changes in opposite-direction or anti-phased.

### 2.3 COMPARISON TO HAAR WAVELET

For easier comparison with the Haar wavelet, we modified it’s mother function as follows:

$$g(z) = \begin{cases} -1, & \text{for } -1 \leq z < 0 \\ 1, & \text{for } 0 \leq z < 1 \\ 0, & \text{for } |z| > 1 \end{cases}$$

where $z=(t-b)/a$, and $a (a>0)$ denotes the scale parameter, $b$ the location/time parameter, and $t$ is time point in the series. The wavelet coefficients for a discrete time series $x(t)$, $(t=1,2,...,N)$, is:

$$W(a, b) = (c_v a)^{-\frac{1}{2}} \sum_{t=1}^{N} x(t) \cdot g((t - b) / a)$$

where the parameters $a>0$ and $b(b=1,2,...,N)$ are integers, and $C_v$ is a constant that
depends on the total power of the Fourier transform of the mother function $g(z)$.

Above formulation represents that the coefficient $W(a,b)$, in (5), of the Haar wavelet (4) is a weighted difference of subseries means between two subseries, which are divided at $b$ with a same sub-sample size $a$, that is, a linear contrast between the two subseries. Other discrete wavelets are formed by modifying the mother function $g(z)$. An abrupt change of the first moment occurs as an enough large absolute value of the wavelet coefficient between two adjoining sub-series at point $b$ on scale $a$, that is, as a large absolute difference in the weighted subseries-means, but unfortunately without means of telling when a difference is large enough to be statistically significance. As an example, the Haar wavelet coefficients $W(a,b)$ for the raw precipitation reconstruction series (Fig. 2.2) produce a pattern very similar to that from the scanning $t$-test (Fig. 4.2a), except for no basis for determining when the values represent a statistically significant change in the subseries-means.

Comparing the statistic $t(n,j)$ in (1) to the Haar wavelet $W(a,b)$ in (5), the sub-sample size $n$ is comparable to the wavelet scale parameter $a$, the reference time point $j$ corresponds to the wavelet location parameter $b$. Both examine linear contrasts of two adjacent sub-series at the time point $j$ or $b$. Figure 2.2 shows the pattern similarity of the two results, detail analysis will be described in §4.3.1. The main differences between the scanning $t$-test and the Haar wavelet are as follows:

(1) The scanning $t$ test gives a test of the statistical significance for detecting change points, whereas the wavelet did not at that time.

Figure 2.3. Similarity of the scanning $t$-test results $t(n,j)$ (up panel) to the Haar wavelet coefficients $W(a,b)$ (low penal) for the precipitation reconstruction series. Contour intervals are 0.25 and 2.0 respectively, but the contour 0 is hidden. Solid lines denote positive values, dashed lines negative values.
The subsample size \( n \) in the scanning \( t \)-test may vary in sequence from 2 to \( N/2 \), while the scale parameter \( a \) in most implementations of the wavelet transform for discrete series requires in theory, as integer powers of 2. This allows the statistic \( t(n,j) \) to detect abrupt changes on various scales \( (n<N/2) \) at each time point \( j \) like a scanner, while the wavelet transforms produce an orthogonal decomposition of the series so that the different scales are independent. The \( t(n,j) \) statistics are not independent as either \( n \) or \( j \) varies.

The scanning \( t \)-test assumes that the observations are independent to each other (i.e. not auto-correlated) and in the normal distribution. Reasonable corrections are needed if the data do not meet these requirements.

### 2.4 THE SCANNING \( F \)-TEST

Similarly, in order to detect significant changes of subseries variances (the second center moments), the statistic \( F_r(n,j) \) of the scanning \( F \)-test is defined as:

\[
F_r(n,j) = \begin{cases} 
-(S_{j1}^2/S_{j2}^2)/F_\alpha, & \text{for } S_{j2} < S_{j1}, \\
0, & \text{for } S_{j2} = S_{j1} \text{ or } S_{j1} = 0, S_{j2} = 0 \\
(S_{j2}^2/S_{j1}^2)/F_\alpha, & \text{for } S_{j2} > S_{j1}
\end{cases}
\]  

(6)

where the subsample standard deviations \( S_{j1} \) and \( S_{j2} \) are calculated in the same way as in (1), and similarly \( n =2, 3, ..., <N/2, j=n+1, n+2, ..., N-1 \). \( F_\alpha \) is a threshold value on the effective degree of freedom after correction of dependence and in normalized distribution for the time series. A local minimum in \( F_r(n,j)<-1.0 \) denotes a significant change towards a smaller variance, i.e. the records become much steadier, whereas a local maximum in \( F_r(n,j)>1.0 \) indicates a significant change towards a larger variance, i.e. the records become much unsteadier. In this algorithm, the subseries variance measures deviations from the subseries mean. So far, no similar mother function has been reported in the wavelets analysis for detecting significant changes of subseries variances.

The estimation of effective degree of freedom for correcting the dependence is taken as (Hammersley, 1946):

\[
Ef(n) = f(n) \cdot \left[ \sum_{\tau=0}^{k} r^2(\tau) \right]^{-1}, r(k) \rightarrow 0,
\]

(7)

where \( f(n) \) is the degree of freedom listed in the \( F \) form.

### 2.5 ALGORITHM FOR THE SERIES WITH UNEQUAL INTERVALS

For those series, for instance, the sedimentary records in (see §4.4) with unequal sampling intervals, we let:

1. the reference point \( j \) denote the order of records in time sequence, while let \( Y(j) \) indicate the corresponding year at \( j^{th} \) record in the time series.

2. The subsample size \( n \) is taken as \{\((j-1)-(j-n)\)\} or \{\((j+n-1)-j\)\}, while the time scale (subperiod duration) is accounted by \{\(Y(j-1)-Y(j-n)\)\} or \{\(Y(j+n-1)-Y(j)\)\} correspondingly, which equalizes approximately the given scale.
In section §4.4, the sample size (number) \( n \) is usually little less than that the time scale is. So the year-error is usually less than that corresponding record interval.

§3 APPLICATION TO FINISH TREE-RING CHRONOLOGY

3.1 INTRODUCTION

Dendroclimatology is the study of tree-ring growth in the context of climate variability. In Finland, dendroclimatology may recall back to Laitakari [1920] who studied the tree-rings of Scots pine and their climate associations in southern part of the country. Sirén [1961] constructed a near-millennial Scots pine ring-width chronology for the northern Finland. This study could be seen as a starting point of constructing long tree-ring chronologies in the region. Later, Briffa, et al. [1990] reported a 1,400-year tree-ring record related to summer temperatures in the northern Fennoscandia. Lindholm et al. [1999] presented a ring-width chronology from northern Finland for the past two millennia. Later, Eronen et al. [2002] completed the supra-long chronology for the past 7500 years. Using the data of this chronology, Helama et al. [2002] reconstructed the regional July temperature variability for the same interval. That study was based on the regional curve standardization (RCS) method and the stepwise multiple linear regression, respectively, used to remove the non-climatic tree-ring variations and to build the transfer function model for the reconstruction. Helama et al. [2002] also described the warmest and coldest periods on intervals of 10, 30, 100 and 1000 years based on their data. Later, Ogurtsov et al. [2005] analyzed periodicities on multicentennial (250–450 yr) and century-scale (90–130 yr) by using Fourier spectrum and wavelet approaches.

In this section, we attempt to detect significant changes of subseries means in this long tree-ring chronology for the past 7638 years [Helama et al. 2008a] by applying the scanning \( t \)-test as shown in §2.2. Here, we describe the data procession for smoothing the high frequency fluctuations by first using Gaussian filter of 81 years, which follows the Maximum Entropy Spectrum analysis; then demonstrates the results of the scanning \( t \)-test, as both consecutive figures and tables for the 21 episodes of subseries means on multi-centennial scales; following by discussing the verification and comparison with glaciological evidence as previously presented for the Greenland ice-cores [O’Brien et al. 1995] and for the Holocene advances and retreats of the glaciers in the Alps [Joerin et al. 2006]. Finally, give a summary for this section.

3.2 DATA PROCESSION

The main data of this study comprise the supra-long tree-ring index chronology of Scots pine (\textit{Pinus sylvestris} L.) as built in the northern Finnish Lapland for 7638 years (here, 5633BC-2004AD). The portion of this data roughly for the past 7500 years (more precisely, 5520BC-1992AD) was first described by Eronen et al. [2002], with a slight temporal extension of the chronology described by Helama et al. [2008a]. In this section, the individual tree-ring series of the chronology were first detrended using a 128-year spline function. The tree-ring index values of individual samples were obtained as division
between the observed tree-ring width and the spline curve. These index values were averaged for each calendar year (5633BC-2004AD) using an arithmetic mean to build the chronology of the tree-ring index to be used for the subsequent analyses.

The distribution of the annual tree-ring indices is very close to the Normal Distribution (Figure 3.1). To smooth high frequency variation before performing the scanning t-test, we adopt firstly the Maximum Entropy Spectrum analysis to the tree-ring chronology and reveal two significant periodicities, there occurring on the 80.3-87.7-year and 30.8-31.8-year scale, respectively (Fig.3.2). Of note, these periodicities are similar to those detected previously from tree-ring data in the same region [e.g. Briffa & Schweingruber 1992]. Then we adopt the Gaussian low-pass filter of 81 years to the index records and produced a 'low-pass' series of the chronology, which will be then used to detect the significant changes on longer scales, by using the scanning t-test as demonstrated in the next section.

The reasons of applying the scanning t-test to a 'low-pass' series can be seen in the light of the data structure, which can be exemplifying as follows. Though the tree-ring chronology is usually serially non-independent (i.e., autocorrelated) and this is also the case of the present data [e.g. Helama et al., 2009]. Moreover, the low-pass filtering makes the data even more dependent in time. However, the extent of autocorrelation becomes temporally relatively homogenous for the full length of chronology, once the same low-pass filter of 81 years is implemented, the dependent autocorrelation is taken correction following the Table-Look-Up Test [Von Storch & Zwiers, 1999] in our algorithm. Moreover, the detection of significant change-points by the scanning t-test is based on picking up the local maximum or minimum centers/points in the $t_{r}(n,j)$ by comparing those at neighbor grids. It follows that it is reasonable to apply the scanning t-test to the low-pass filtered chronology as a reference tool of detecting change-points.

![Figure 3.1 The Maximum Entropy Spectrum analysis of the tree-ring index in Finnish Lapland for 7638 years](image)
3.3 RESULTS

Figure 3.3a presents the contours of the results $t_{r(n,j)}$ for the tree-ring chronology of the Finnish Lapland over the past 7634 years. The centers in positive values of the $t_{r(n,j)}$ denote uplifting changes of subseries means, while centers in negative $t_{r(n,j)}$ indicate descendant changes. This is exemplified by the earliest significant change-point which is found as a center in negative (decline change) in 5258BC on time-scale of 140 years (see also Fig. 3.3b, the first vertical thick line) with the following episode mean of -0.050 (see also Fig. 3.3b, the second flat thick line) and variance of 0.863 (see also Fig. 3.3b, the second flat thick dashed line) for 5258BC-5050BC. This episode mean is lower than the -0.004, as compared with the earliest episode during the interval of 5633BC-5259BC. The second change-point appears as a center in positive, as demonstrated by an uplifting change in 5049BC on time-scale of 180 years, with the following episode mean +0.084 and variance value 0.916 for 5049BC-4719BC. The third change-point is a negative center in 4718BC on time-scale of 304 years with the episode mean of -0.037 and variance of 1.118 for the period of 4718BC-4351BC. Similarly, the following change-points and corresponding episode-means are plotted in Fig. 3.3b as the thick line. The episode variances are drawn as the thick dashed line.

Figure 3.2 a) The contours of the $t_{r(n,j)}$ for the tree-ring dendrochronology of Finnish Lapland for the past 7638-years. In x-axis, the negative values are years in BC, the positive values are years in AD. b) the low-pass filtered tree-ring chronology (dashed thin curve), the change points and the episode means (thick solid line), and the episode variance (thick dashed line)
The successive negative centers on the top of contours for the period between 4960 BC and 2442 BC imply a decreasing tree-ring level of long term change (indicating palaeoclimatic cooling), while the large positive area between the years 2441 BC and 449 BC denotes an increase tree-ring level (indicating palaeoclimatic warming). A roughly normal stage, with a normal palaeoclimate, occurred between 425 BC and 1148 AD, as demonstrated by the thick line in Fig. 3.3b. Over the late Holocene, the last 850 years appear to contain three episodes of obvious changes in the tree-ring proxy: the episode 1149-1447 AD showing notably high tree-ring indexes with episode mean of 0.086; the episode 1448 AD-1735 AD with low tree-ring indexes at episode mean of -0.105; and the most recent episode 1736-2004 AD, which demonstrates again higher index values with episode mean of 0.063.

Table 3.1. Change-year, episode duration, episode mean and variance, and verifications

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<tr>
<td>20th</td>
<td>1736 ↑</td>
<td>268</td>
<td>+0.063, H</td>
<td>0.949, u</td>
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<td>CT30</td>
<td>C</td>
<td>Advan.</td>
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<td>+0.086, SH</td>
<td>0.957, s</td>
<td>s</td>
<td>W</td>
<td></td>
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<tr>
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<td>Advan.</td>
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<td>431</td>
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<td>CT30</td>
<td>Cs</td>
<td>Advan.</td>
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<td>Reces.</td>
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<td>C</td>
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<td>0.736, s</td>
<td>s</td>
<td>Ws</td>
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<td>Reces.</td>
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<td>Cs</td>
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<td>375</td>
<td>-0.004, N</td>
<td>1.197, u</td>
<td>s</td>
<td>W</td>
<td>Reces.</td>
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Table 3.1 lists the order of 20 change-points in the first column, and corresponding change years, with the arrowheads indicating the change directions in the second column. The third column lists the episode durations following the change years. The durations
varying from 105 to 845 years, with average of 363.7 years. The fourth column depicts the episode mean in the left, and the right-hand side giving the grade corresponding to the episode-mean in 5 grades as the SH (Sever High)>0.085, 0.085≥H(High)>0.025, 0.025≥N(Normal)>-0.025, -0.025≥L(Low)>-0.059 and -0.059≥SL(Sever Low). Statistically, the episode mean in the Normal grade occupies a total of 2616 years in 34.2%, the High and Sever High grades have 2194 years in 28.7%, the Low and Sever Low grades are total 2827 years in 37.0%, which is more than that in the High +Sever High grades.

The fifth column lists episode variance values in left, the right letters “s” means stable, denoting the variance value <1.00, and “u” means unstable, corresponding to the variance value ≥1.000. In statistics, there are altogether five stable episodes, with comparison to only two unstable episode variances in the H (High) grade plus SH (Sever High) grade of episode means; 3 stable and 2 unstable of variances in the N (Normal) grade; while 3 stable relating to 6 unstable of variances in the L (Low) and SL (Sever Low) grades. The third columns from the sixth to eighth present verifications and compatibilities with previous published papers. These will be discussed in detail in the next section.

3.4 DISCUSSION

It should be noted that our findings are somewhat different from the findings of Cook and Peters [1997]. They concluded that the factors inherent to tree growth, not climate, lead to higher mean (that is, generally wider rings) can be associated with the high variance (that is, pronounced alternation of wide and narrow rings). The observed dissociation between the mean and variance in our results (Table 3.1) should be seen as a result of successful tree-ring standardization that was applied to the tree-ring data before any climatic analyses. That is, the standardization is known to largely remove such heteroscedasticity from the tree-ring data [Fritts 1976; Cook & Peters 1997; Helama et al. 2004]. Once the characteristics of the tree-ring chronology have been demonstrated using the applied methodology, the described results (e.g. the change years and episodes in Table 3.1) can be compared with the results and palaeoclimatic interpretations from previous studies. This is done in the context of the previously published dendroclimatic [Helama et al. 2002] and glaciological evidence from the Greenland ice-cores [O’Brien et al. 1995] and the Apls [Joerin et al. 2006].

3.4.1 COMPARISON WITH A SIMPLE METHOD

Previously, the extreme tree-ring episodes have been analyzed using simple mean of all possible 100-year periods for the practically same tree-ring material [Helama et al. 2002] as cited here. The coldest such period was seen having occurred during the period of 5208BC-5109BC. We note that this period agrees with the first of our episodes having occurred during the period of 5258BC-5050BC, so it is listed here as “CT100” (the coldest 100-years) as reference in the fifth column of Table 3.1. Likewise, the warmest of the 100-years as found in a previous study was evident for the period of 3481BC-3742BC [Helama et al. 2002], this anomaly coinciding our finding of high tree-ring indexes during the episode between the years 4002BC and 3632BC (see “WT100”, the warmest 100-years, referred in Table 3.1). Moreover, palaeoclimatologically the coldest 100-year intervals, detected for the period of 2500BC-2401BC [Helama et al. 2002] was highly
compatible with our episode as found for the period of 2567BC-2391BC ("CT100" in Table 3.1). Yet, we found that the warm 100-year interval between the years 2400BC and 2301BC in the previous study [Helama et al. 2002] is contained in the episode for the period of 2390BC-1763BC ("W100" in Table 3.1). Similarly, the warmest 100-year periods during the intervals of 2622-2523BC, 1238-1139BC, 604-505BC and 28AD-127AD, appear to correspond the episodes we have detected for the periods of 3412-2568BC, 1538-998BC, 704-426BC and 6AD-530AD respectively. Interestingly, the most recent warm intervals, as found previously for the period of 1901AD-1992AD [Helama et al. 2002] and here, as an episode of 1736AD-2004AD, appear to contain the similar signal of warming and thus more positive tree-ring index values (Table 3.1). With regards to the variations of shorter duration, our 7th episode occurring between the years 3631BC and 3413BC, includes the coldest palaeoclimatic 10-year period as previously detected for the period 3453BC-3444BC [Helama et al. 2002]. Moreover, the coldest 30-year intervals as detected in that study, the periods of 1648-1619BC, 422-393BC, 542AD-571AD and 1459AD-1488AD, largely corresponded with the episodes in our Table 3.1 for the periods of 1762-1539BC, 425BC-6AD, 531AD-1148AD and 1448AD-1735AD respectively.

Totally, 14 of our 21 episodes were found being compatible with the previous dendroclimatic results [Helama et al. 2002]. Notably, no correspondence was for the remaining seven episodes. Likely, this reflects the differences in the way the data was analyzed, with the previous focus on predominantly the warmest and coldest intervals during each consecutive millennia. With regards to our earliest episode as found for the period of 5633BC-5259BC, as well as the most recent episode as found for the period 1736AD-2004AD, it might be that the edges of the time series have influenced the analyses differently.

Although majority of the episodes, as detected in a previous study [Helama et al. 2002], using a simple method, and here, using a more advanced algorithm, were overlapping, the method presented in the study in hand may be seen notably advantageous. In a method, the episodes of extreme palaeoclimatic circumstances were determined merely by selecting the most positively and negatively anomalous 10- and 100-year periods from unweighted running averages for each millennia. First, the method may be insensitive for such periods spanning over the turn of two consecutive millennia. Second, the simpler method was unable to detect two or more anomalous periods for each millennium, whereas the new method can detect several anomalous episodes per millennium (see Table 3.1). Third, the period length of 10 or 100 years may be seen arbitrary, as used in the simpler method, whereas the detection of the tree-ring episodes by the new method is not limited by any fixed period length (see Table 3.1) being thus far more data-adaptive method in character. Fourth, the new method defines a continuous track of tree-ring episodes, not merely a list of periods with most anomalous palaeoclimatic conditions. Fifth, the statistical significance of the significant changes of subseries means and variances are determined. By contrast, the simples method of episode detection [Helama et al. 2002] did not contain statistical testing of the palaeoclimate episodes and their mean levels. We see all these five issue as notable advantages from a rather primitive calculation of the most anomalous mean levels over more or less arbitrary pre-selected period lengths.
3.4.2 COMPARISON WITH GLACIOLOGICAL EVIDENCE FROM GREENLAND

The significant changes in tree ring chronology may be influenced by other factors besides climate change, for instance, the changes of tree ring samples, which are usually collected from different geographic sites. It is needed to compare with other evidence to identify if the change point in the tree ring chronology indicates a climate change really.

The Holocene climate was synthesized previously in a benchmark paper of O’Brien et al. [1995]. Their study was based on the reconstruction of the EOF1 form 5114 samples retrieved from the Greenland ice sheet project 2 (GISP2). The EOF1 component was referred to as the pole circulation index [O’Brien et al. 1995]. Their main findings include the noticeable increases in the Holocene EOF1 values, which associates the northern polar vortex expansion or stronger meridional circulation and colder climate, as having occurred during the periods of 0-610BP and 5000-6100BP, and earlier than 11300 BP (the years given here following the method of the original paper, BP indicating the years before 1950AD). Moreover, increases of lesser magnitude occurred at 2400-3100BP and 7800-8800BP, while the milder climate occurred at years about 610-960BP, 1500-2700BP, 6300-7900BP, and 9300-10600BP [O’Brien et al. 1995]. In addition to these major anomalies, it is noteworthy further details of the EOF1 curve [see Fig.2 in O’Brien et al. 1995] indicate that cold climates of shorter duration may have taken place during the period of 7250-7100BP, 6500-6300BP, 4000-3450BP and 2200-1900BP, while similarly warm climates may have prevailed in Greenland during the period of 6300-6100BP, 5000-4100BP, 2700-2200BP and 1900-960BP. We note several interesting similarities between their [O’Brien et al. 1995] and our results (Table 3.1). These are exemplified by tree-ring episodes during the periods of 5633-5259BC, 5049-4719BC, 704-426BC, 6AD-530AD and 1149AD-1447AD, broadly coinciding with the Greenland mild climate phases during the periods of 6300-7900BP, 1500-2700BP and 610-960BP (marked as "W" in Table 3.1). Moreover, tree-ring episodes of 4245-4003BC, 3631-3413BC and 1448AD-1735AD overlap with the Greenland cold climate phases as recorded for the periods of 5000-6000BP and 0-600BP. Similarly, our episode 997-705BC overlaps with the Greenland cold climatic phase during the period of 2400-3100BP ("C" in Table 3.1). Yet, the tree-ring episodes of 5258-5050BC, 4718-4351BC, 1762-1539BC and 425BC-5AD are compatible to the Greenland cold climate phases of shorter duration during the intervals of 7250-7100BP, 6500-6300BP, 4000-3450BP and 2200-1900BP ("Cs" in Table 3.1). It is also notable that the tree-ring episodes during the periods of 4350-4246BC, 3412-2568BC, 2390-1763BC and 1538-998BC agreed with the Greenland warm climates of shorter duration as recorded for the periods of 6300-6100BP, 5000-4100BP, 2700-2200BP and 1900-960BP ("Ws" in Table 3.1). In summary, 17 of 21 tree-ring episodes were compatible with the Greenland climate changes as recorded in the EOF1 variations of the ice-core data [O’Brien et al. 1995].

It is interesting to note that the climate conditions in Greenland and Europe are often seen to present a see-saw behavior, where the milder climate in Greenland appears coeval to cooler circumstances in Europe, and vice versa [e.g. Olsen et al. 2012]. Physical reason for this see-saw is the atmospheric patterns known as the North Atlantic Oscillation (NAO). The positive phase of NAO come with strengthened (attenuated) westerlies and thus milder (cooler) climates, especially in winters [Hurrell 1995; Hurrell et a. 2001]. However, the NAO indexes correlate positively in the study region also during the summer...
months [Helama et al. 2008b]. This means that the NAO could be relevant for interpretations of the tree-ring data of the present study. By contrast, our comparison revealed several periods during which the palaeoclimate conditions were of similar character (Table 3.1), not reversed as could be expected from the see-saw behavior of the NAO. Consequently, the NAO pattern could not be seen responsible for these climate anomalies. This interpretation follows the finding that similar phases (e.g. synchronous cooling in both regions) of climate extremes can take place in Greenland and Nordic [Nesje & Dahl 2001]. Here we have detailed such periods for the past seven and half millennia (Table 3.1). Potentially, our findings are to pave a way for an advanced future interpretation of natural climate variability around the North Atlantic region.

3.4.3 COMPARISON WITH GLACIOLOGICAL EVIDENCE FROM ALPS

Records of mountain glacier lengths provide palaeoclimatic information of past climatic changes [Oerlemans 2005]. In the Swiss Alps, Joerin et al. [2006] summarized major periods of glacier advances and retreats, based on 143 dated wood and peat fragments. Their dating results were given in calibrated years before 1950AD and rounded to the nearest 50 years. Transferring these dates into the calendric AD and BC years, their major periods of the glacier recessions (indicating warming climate) can be listed as 5750-5600BC, 5500-4600BC, 4200-4000BC, 3750-3550BC, 3250-2450BC, 2350-1450BC, 850BC-750BC, 200BC-100AD and 550AD-750AD for the last 7700 years. We note that at least four of the positive tree-ring episodes may well be comparable with the glacier recessions in the Alps. That is, the tree-ring episode during the period of 5633-5259BC may link to the late part of the glacier recession during the period of 5750BC-5600BC, and the glacier recession 5500-4600BC overlaps with our tree-ring episodes during the years 5049-4719BC. Moreover, the tree-ring episode over the period 3412-2568BC overlaps mainly with the glacier recession in the Alps occurring 3250-2450BC, whereas the tree-ring episode of the period 2390-1763BC is broadly synchronized with the glacier recession during the period 2350-1450BC (marked as “Reces.” In Table 3.1).

According to Joerin et al. [2006], the interruptions between the abovementioned glacier recessions can be considered as periods of glacier advances, indicating climatic cooling. Comparison with our results shows that the tree-ring episode during the period of 4718-4351BC coincide with the glacier advance between the years 4600BC and 4200BC, and that the tree-ring episode during the period of 2567-2391BC may well be comparable with the glacier advance between the years 2450BC and 2350BC. Moreover, our tree-ring episode that spanned the period 425BC-5AD stared during the glacier advance after 750BC but before 200BC. Yet, the tree-ring episodes occurring during the period 531AD-1148AD and 1448-1735AD may be seen to overlap with the long glacier advance in the Alps that started around the year of 750AD (“Advan.” In Table 3.1). In conclusion, 9 of 21 tree-ring episodes in northern Finland (Lapland) appear coinciding the glacier episodes in the Alps (see Table 3.1).

Besides above cited, first, our findings relate to the results of Karlén [1991], who made an encouraging comparison of his glacier record, drawn from historical records, radiocarbon dating of wood, lichenometric dates on moraines, with his local tree-ring evidence over the last 15 centuries. Second, over broader spatial scale relevant to our
study, the findings shown here (Table 3.1) are also following those of Briffa et al. [1999], who showed that the European climate north of the Alps cooled synchronously during the last decades of the 16th century AD. Third, a more recent study [Helama et al. 2010] showed that the major phases of glacier advances, as recorded by Denton and Karlén [1973], were broadly synchronous with the tree-ring based climate reconstruction from the study region over the present inter-glacial era [Helama et al. 2010]. Here we are adding into these findings by detailing the synchronicity of the palaeoclimatic information from tree-rings and mountain glacier movements in Europe. It is also noteworthy that unlike the previous studies, which had found correlations between climatic cooling and glacier size advances, our new results (Table 3.1) showed not only four periods with indications of cooler climates but also five periods with warming signals. That is, the results shown here demonstrated the tree-ring correlation with glacier retreat phases, thus crucially adding for the information of previous investigation. Detailed investigations will be needed to find out the physical mechanisms in the forcing-climate dynamics responsible for these variations. Likely, such factors include the influence of the Sun, in addition to variations in the internal climate modes and even the volcanic impacts [Nussbaumer et al. 2011].

3.5 SUMMARY
This section presents significant changes in the subperiod means by applying the scanning t-test algorithm to the low-pass filtered series of the Scots pine tree-ring chronology in Finnish Lapland for the past 7638 years. The low-pass filtered chronology was produced by 81-year Gaussian filter. The selection of this filter was judged by the findings of significant periodicities of 80.3-87.7 years and of 30.8-31.8 years, as found in the tree-ring chronology, by using the Maximum Entropy Spectrum analysis.

Twenty change-points and 21 episodes were found and showed with characteristically high and low episode-means by using the scanning t-test. The durations of the episodes occurred on multicentury time-scale, ranging from 105 to 845 years, with an average of 363.7 years. The episodes were classified into five grades according to their mean tree-ring index values. For longer time scales, there appeared a decreasing tree-ring trend over the period 4960BC-2442BC, and a positive trend from 2441BC till 449BC. These trends were followed by a roughly normal stage between the years 425BC and 1148AD. Three episodes in the last 850 years occurred as obvious changes: the episode 1149-1447AD in sever high, the episode 1448AD-1735AD in sever low, the last episode 1736-2004AD changing into the high index level, again.

All 21 episodes were verified by comparing with one or more previously published papers on the records of past climate variability. It was found that 15 episodes were verified by two or more previous studies, six episodes were referred to at least one of the previous works. The change-points, episode features including the mean and variance, and the verification references are all summarized in Table 3.1. [Timonen, Jiang, et al, 2014]

§4 APPLICATION TO USA PRECIPITATION RECONSTRUCTION

4.1 INTRODUCTION
Hughes & Graumlich [1996] presented a valuable dendroclimatic study of annual (prior
July through current June) precipitation reconstruction for the period from 6000BC to 1997AD for climate division 3 in south central Nevada, based on the Methuselah Walk tree-ring chronology from the White Mountains in California, the southwestern USA (Fig. 4.1). They emphasized the striking pattern of two multi-decadal droughts in the epoch between 900AD and 1400AD and listed eight extreme drought years, in which the bidecadally filtered values of the precipitation reconstruction were below 17 cm. However, no analysis of statistic significant changes was reported for this series before.

This section attempts to apply both algorithms of the scanning $t$-test [Jiang et al. 2001, 2002] and the scanning $F$-test [Jiang et al. 2003], as mentioned in §2.2 and §2.3 respectively, to the unfiltered precipitation reconstruction for Nevada climate division 3, here the unfiltered precipitation reconstruction represents that has not been filtered bidecadally [Hughes & Graumlich, 1996]. Firstly, the scanning $t$-test identifies twenty two change points, and 23 comparatively Dry/Normal/Wet climatic wetness-episodes are partitioned in the 8000-year precipitation reconstruction series. Secondly, the scanning $F$-test detects 15 change points of subperiod variance and divides 16 phases in comparatively steady (with smaller variance) or unsteady (with larger variance) features. Thus the 23 wetness episodes are characterized as the steady or unsteady traits by jointing the results from the scanning $F$-test into those of the scanning $t$-test. Thirdly, in order to verify the significant changes of subperiod levels in the unfiltered precipitation reconstruction series, we employ the coherency detection algorithm based on the scanning $t$-test [Jiang et al. 2002] to the precipitation reconstruction and other two high-resolution sediment time series, the TIC (Total Inorganic Carbon fraction) and $\delta^{18}O$ records from cored sediments in deep basin of the Pyramid Lake (see Fig.4.1) in Nevada [Benson, et al., 2002]. The coherency detection algorithm tests synchronously and asynchronously significant changes in the subperiod levels between two time series on multi-time scales. There appear thirteen change points in the precipitation reconstruction
nearing to those in the TIC or $\delta^{18}$O series within 150 years of difference and covering over-lapped years more than 2/3 of the episode duration. Finally, we confirm the 23 wetness-episodes with previously published researches into the climatic change periods in the western USA, and find that 22 of the 23 episodes are coincided with the previously published results. In addition, the 23 episodes are also compared with studies of climate changes in the eastern China as well as of the global changes. As known, the eastern China and western USA locate western and eastern coasts of the Northern Pacific respectively, and China locates in the upstream, USA locates in the down-stream of the westerlies in general atmospheric circulation. They are all influenced by the ENSO, and PDO and other factors but with different aftereffects.

4.2 Data Sources

Three time series, the unfiltered precipitation reconstruction, the TIC and $\delta^{18}$O records, are used in this work. The unfiltered precipitation reconstruction series was downloaded from the website [1], [Hughes, M.K. and L.J. Graumlich, 2000, Multi-Millenial Nevada Precipitation Reconstruction. International Tree-Ring Data Bank. IGBP PAGES / World Data Center-A, for Paleoclimatology Data Contribution Series #2000-049. NOAA / NGDC Paleoeclimatology Program, Boulder CO, USA]. The series contains a reconstructed annual (prior July-current June) precipitation (in cm) for Nevada Climate Division 3 (see the NVCD3 in Fig.4.1) including a total of 7997 years from 6000BC to 1996AD (referred to as the precipitation reconstruction hereafter). It was reconstructed by using the whitened Methuselah Walk tree-ring chronology from the White Mountains of California, the location is shown as number ‘0’ in Figure 4.1. Other numbered sites in Figure 4.1 will be mentioned below and in subsection 4.5. This tree-ring chronology is the longest absolutely dated in a single species of the bristlecone pine, at almost 9,000 years long. It is made up of 285 tree-ring samples with average length of 748 years, of which about 14 series are present in each year for the whole period after 6000 BC. These tree-ring samples were collected by the Laboratory of Tree-ring Research, University of Arizona. In order to remove any growth trend of the trees, they standardized all tree-ring samples with either negative exponential or a straight line with zero or negative slope. The Box-Jenkins model was fitted (ARMA 1, 1), which accounted for only 5.74% of series variance, and the series was whitened by computing the residuals from this model. The whitened Methuselah Walk chronology was calibrated in regression to annual precipitation observations in the Nevada Climate Division 3 during the period from 1932 to 1979, and then extrapolated to other years. This regression fitted 35% variance of the precipitation observations. Both power spectrum and singular spectrum analyses indicate that very little of variance of this chronology can be represented by trend, or periodic or quasi-period components. More details about the data are described in Hughes & Graumlich [1996].

The TIC and $\delta^{18}$O high-resolution records from cored sediments in deep basin of the Pyramid Lake (see number “1” in Fig. 4.1) in Nevada were kindly provided by Dr. Benson at USGS. These data sets were sampled from 2 piston cores of sediments in the center Pyramid Lake in 1997 (referred as PLC97-1) and in 1998 (PLC98-4) separately. The sites
of these 2 cores are very close to each other. The PLC97-1 covers the period from 793BC to 1839AD and 533 records were read at unequal intervals between 3 and 9 years. Its age control was established by comparing its paleomagnetic secular variation (PSV) record with a well-dated western USA archeomagnetic record (Lund, 1996), and the age model accuracy was considered within 50-100 yr. The PLC98-4 was taken to recover older Holocene-age sediments, which covers the period from 5680BC to 1480BC, and 538 records were read in unequal intervals between 4 and 14 years. Radiocarbon ages were determined on the TOC (Total Organic Carbon fraction) of this core and probably remained larger errors. Further descriptions of the TIC and δ¹⁸O data sets can be found in Benson et al. [2002].

4.3 CHANGEPOINTS IN THE PRECIPITATION RECONSTRUCTION SERIES

4.3.1 RESULTS FROM THE SCANNING t-TEST

The scanning t-test was computed for time scales (subsample sizes) ranging from 54 to 2896 years following equation (1) and (2) at 95% confidence level. We calculate the time scales up to 2896 years in consideration mainly of two reasons: one is the mathematical capability, another reason is that more than 10 tree-ring series cover the whole period after 6000 BC. Based on the local maximum and minimum of the t-test values, twelve positive (increases in precipitation reconstruction) and ten negative (decreases in precipitation reconstruction) centers were picked up, i.e. 22 significant changepoints were discovered by our computing program (Fig. 4.2a).

Fig. 4.2  (a) Contours of the normalized scanning t-test for the precipitation reconstruction series at confidence level 95%. Contour interval is 0.25 but the zero-contour is hidden. Solid lines denote positive values, dashed lines negative values.  (b) Change points and episode averages (solid line) from Fig.4.2a, 101-year Gaussian filtering of the precipitation reconstruction (dashed line).
For example, the first significant change point towards increase in precipitation reconstruction is detected with a positive center of the contours around 5865BC on 128-year time scale. It was followed by a negative center, a decrease in precipitation, in about 5706BC on a little shorter time scale. Then the precipitation reconstruction increased once more in 5339BC on 91-year scale and in 4606BC on 304-year scale respectively. The second decrease point is in 4313BC on time scale 362 years. Then two significant increases occurred in 3770BC and 3205BC on time scales 724 years. These may reveal a step-wise approximation to what appears to have been a longer period of increasing precipitation in the series. A few local centers of the contours on time scales longer than 750 years, such as in 3400BC, possibly result from those tree-ring series covering the whole period after 6000 BC. Every change point is shown in Fig. 4.2b and listed in Table 4.1.

Table 4.1. Change points and Wet /Normal/Dry episodes in the Precipitation reconstruction series, and sources of collaborative evidence (TIC, δ¹⁸O, paleoclimate references) for each episode.

<table>
<thead>
<tr>
<th>Change-years (AD)</th>
<th>Episode span, &amp; grads</th>
<th>TIC Change years</th>
<th>δ¹⁸O change years</th>
<th>Quoted references</th>
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<td></td>
<td></td>
<td></td>
<td></td>
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<td>(1839)</td>
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<td>1606↑</td>
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<td></td>
<td>1477↓</td>
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<tr>
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<td>141, W</td>
<td>794↓</td>
<td>768↑</td>
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<tr>
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<td>452, D</td>
<td>538↑</td>
<td>529↑</td>
<td>[19]</td>
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<tr>
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<td>227↓</td>
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*Notes: The sign minus denotes years in BC.

↑: indicates a change to wet; ↓: change to dry.
Numbers in the second column denote the span from that change year to the next.
SD: Severe Dry; D: Dry; N: Normal; W: Wet; SW: Severe Wet.
( ) : The beginning and ending years of the data.


These 22 change points partition the unfiltered precipitation reconstruction series into 23 episodes of relatively dry or wet subperiods, which are estimated from the episode levels of precipitation reconstruction for that episode (Fig. 4.2b). The episode averages can be sorted into 5 grades of climate-scale wetness: Severely Dry (SD) < 19.1 cm; 19.1 < Dry (D) < 19.6 cm; 19.6 < Normal (N) < 19.9 cm; 19.9 < Wet (W) < 20.3 cm and Severely Wet (SW) > 20.3 cm, respectively (Table 4.1). It suggests that this region is in an arid-semiarid climatic category. Compared to Hughes & Graumlich [1996], who analyzed the bidecadally filtered precipitation reconstruction series, the eight years labeled as extremely droughts in their Table 2 are ranked as Severe Dry or Dry episodes in our partition. The striking multi-decadal droughts between 900AD and 1400AD in Hughes & Graumlich [1996] are very similar to the Dry episode 898-1319AD in our Table 4.1.

The 23 episode durations span from 131 years (the SW-episode in 1710AD-1840AD) to 736 years (the SW-episode in 278BC-457AD) with average of 348 years (Table 4.1). It suggests that the precipitation changes on multi-centennial time scales in this analysis. This result might be technologically reasonable in the dentochronology, because the longest episode-duration 736 years is shorter than the average length 748 years of the tree-ring samples. Among the 5 grads, the SD and D-grads take over totally 2845 years, around 35.6% of the total 7997 years, the N-grad occupies sum of 2664 years, 33.3%, and the W-SW grads cover 2487 years in sum, i.e. 31.1% of the total 7997 years.

4.3.2 RESULTS FROM THE SCANNING F-TEST
The scanning F-test (equation 4) of the precipitation reconstruction series was calculated on same time scales as in Figure 4.2. Figure 4.3a shows many frequent variations on short time scales. On time scales longer than 128 years, however, seven positive (increases of subseries variances) and eight negative (decreases of subseries variances) significant changes were detected with local maxima and minima centers in the contours (Fig. 4.3a). The change years are usually different from those in the first moment (subperiod levels). It is featured generally smaller standard deviation, which means a steadier climate, in latter period after 2000BC than that in the earlier period. We may characterize each episode as a steady or unsteady feature by combining these results with those episodes that were partitioned in section 4.3.1. For instance, the episode 5339BC-4607BC is featured as unsteady (symbolized as “u” in the second column of Table 4.1), because a large standard deviation (Fig.4.3b) corresponds to it. The feature steady (symbolized as “s” in the second column of Table 4.1) for the episodes 4606BC-4314BC and 4313BC-3771BC follow, because smaller standard deviation (Fig.4.3b) coincides with them comparatively, and so on.

Statistically from Table 4.1, there are 8 steady in proportion to 1 unsteady of total 9
episodes in the SD-D grads, 5 steady in ratio to 3 unsteady of 8 episodes in the W-SW grads, while 2 steady comparing to 4 unsteady of 6 episodes in the N-grad. Why the W-SW grads have slightly higher ratio in unsteady episodes than those in the SD-D grads, perhaps are partly influenced by the tree growth, which is not perfectly removed in the reconstruction process, because it leads to that the high mean (increased tree ring growth or width) is naturally related to high variance (increased growth variance) by factors that are internal to tree growth, not to climate [Cook and Peters, 1997]. However, five steady of eight high-mean episodes W-SW are still dominant. This implies here the $F$-test of the precipitation reconstruction can reflect mainly the significant climate-change information.

Fig. 4.3 (a) Contours of the normalized scanning $F$-test on precipitation reconstruction series at 95% significance level. Contour interval is 0.25 but the zero-contour hidden. Solid lines denote positive values, dashed lines negative values. (b) change points and subperiod standard deviation (thick dashed line) from Fig. 4.3a, change points and episode averages of precipitation (thick solid line) as same as in Fig.4.2b, and unfiltered precipitation reconstruction values (thin dashed line).

4.4. COMPARISON WITH SEDIMENTARY RECORDS

This section presents some approximately synchronous changes of subseries means in the precipitation reconstruction with those in the TIC or $\delta^{18}O$ records, which are built from cored sediments of Pyramid Lake, Nevada [Benson et al. 2002]. Their coherency analyses did not include this subseries-variance changes, because there are multi-millennial trends in both the TIC and $\delta^{18}O$ series, especially in the earlier period from 5680BC to 1480BC (see Fig.4.8). Benson et al. [2002] stated that the lake volume fluctuations and fluctuations
of the $\delta^{18}O$ or TIC records are not simply linearly correlated. In general, droughts lead $\delta^{18}O$ to increase initially as the lake’s volume decreases, but the steady-state value of $\delta^{18}O$ for a long drought period is much smaller than that in a wetter period. During wet episodes, Pyramid Lake receives more inflows from its source, the Truckee River, and thus has higher $\delta^{18}O$ values. TIC is expected to parallel changes in $\delta^{18}O$ for long dry and wet periods.

![Figure 4.4. Contours of the normalized scanning t-test at a 90% significance level:](image)

a) for the TIC series from Pyramid Lake sediments; b) for the precipitation reconstruction series at same year / time-scale grids as in figure 4.4a; c) the coherency of significant changes between a) and b). Contour interval is 0.2 but contour zero hidden. Solid lines denote positive values, dashed lines negative values.

The coherency of significant changes in the precipitation reconstruction series with those in the TIC series for earlier years (Fig. 4.4a) shows four positive coherency (in-phase) centers in 4700BC and 4230BC on 256-year scale, in 3680BC on 2048-year scale, and in 3100BC on 512-year scale respectively, despite the presence of three weak negative coherency centers. Some but not all of the change-points in the individual series (Fig. 4.4a) coincide closely. For instance, the TIC change-points (centers in Fig. 4.4a) in 4704BC, 4236BC, 3646BC and 3114BC are close to those in 4606BC, 4313BC, 3770BC, and 3205BC in the precipitation reconstruction series (centers in Fig. 4.4b or 4.2a) respectively by differences within 150 years. The last weak change around 2803BC in TIC (Fig. 4.4a) preceded that in the precipitation reconstruction series by 158 years, so that the coherency is not obvious, we listed this in Table 4.1 with a question mark. Two possible explanations for the differences of significant change years between the two series are firstly chronological errors, especially in the sediment records comparing to the tree ring reconstruction, and secondly different locations where the data were collected from.
For later years, Figure 4.5 displays 5 positive coherency centers at 200BC, 190AD, 330AD, 640AD and 870AD (comparatively weaker) on scales longer than 300 years, except for some weak negative coherency centers on scales shorter than 300 years. That is, significant changes in the TIC series are mainly in-phase with those in the precipitation reconstruction series on time scales longer than 300 years, while anti-phased on scales shorter than 300 years. This result agrees with Benson et al. [2002] as above mentioned in the first paragraph of this section. The TIC change points (Fig. 4.5a) in 113BC, 538AD, 794AD and 1606AD are approximately close to those in the precipitation reconstruction at 278BC, 652AD, 898AD and 1710AD respectively (Table 4.1 or Fig. 4.8). As the TIC change year 113BC occurred 165 years later than the change point in 278BC in the precipitation reconstruction, we list it in Table 4.1 with a question mark. The TIC change year 214 AD, which preceded the change point 458AD in the precipitation reconstruction by 244 years, is not listed in Table 4.1.
Similarly, Figure 4.6c demonstrates the coherency of significant changes between the $\delta^{18}O$ (Fig. 4.6a) and the precipitation reconstruction (see Fig. 4.6b) for the earlier period, and unmaps five weak positive coherency centers around 4620BC, 3710BC, 3300BC, 2750BC and 2240BC respectively. Why the coherency here is weaker than that between the TIC and the precipitation reconstruction series (Fig. 4.4c), because changes in the $\delta^{18}O$ series itself are weaker. However, the change-points detected in the $\delta^{18}O$ series in 4603BC, 3249BC, 2795BC and 2317BC (Fig. 4.6a), are comparable to those in the precipitation reconstruction series (Table 4.1 or Fig. 4.8). In addition, the $\delta^{18}O$ change-point in 2043BC was 161 years earlier than that in the precipitation reconstruction (Table 4.1 or Fig. 4.8), the coherency index (Fig. 4.6c) is negative in sign, so we list it with a question mark in Table 4.1. Because timescale of the positive coherency center in 3710BC (Fig. 4.6c) is too long to identify changes on multi-centenary timescales in the $\delta^{18}O$ series, no corresponding change year is listed in Table 4.1.
For the later period, there are 6 in-phased changes in 200BC, 230AD, 640AD, 890AD, 1320AD and 1500AD on scales longer than 100 years. The significant changes 277BC, 441AD, 529AD, 768AD, 1308AD and 1477AD in the δ¹⁸O series may correspond to those in the precipitation reconstruction series (Table 4.1 and Fig. 4.8).

Further comparisons, by calculating the episode averages for each series based on the detected change-points for that series and by using a 101-year Gaussian filter to low-pass filter each series, are illustrated in Fig. 4.8. An obvious difference is that the millennial-scale trends in both the TIC and δ¹⁸O series in the earlier period from 5680BC to 1480BC are much more apparent than that in the precipitation reconstruction. As mentioned in section 4.2, the precipitation series is reconstructed from tree-rings with average length shorter than 750 years and with standardization of every tree-ring sample, which may preclude any millennial-scale trends in the precipitation reconstruction series.

Even so, there are common features in the episode averages and smoothed data of the three series. In the middle Holocene, the Dry episode in 5706-5340BC, the Wet episode in 4606-4314BC, the Severely Wet episode in 3205-2646BC and the Dry episode in 2645-2383BC identified in the partitioned precipitation reconstruction series (Fig. 4.8c) have rough equivalents to those in the other two series (Fig. 4.8a and 4.8b). The Dry episode in 4313-3771BC and the Normal episode in 3770-3206BC in the precipitation reconstruction
have analogs in the TIC series, while the Normal episode in 2382-1883BC and 1882-1418BC approximate those in the $\delta^{18}O$ series. Only the Normal episode in 5339-4607BC is not collaborated by either the TIC or $\delta^{18}O$ series.

![Graph of precipitation reconstruction](image)

**Figure 4.8.** The change-points and episode averages (solid line), 101-year Gaussian filter (dashed line) and averages over the entire time series (dotted line) in the three series: (a) for the TIC and (b) for the $\delta^{18}O$ series from sediment cores in Pyramid Lake, (c) as same as in Figure 4.2b;

During the late Holocene, the Severely Dry episode in 784~279BC, the Severely Wet episode in 278BC-457AD, the Dry episode in 458-651AD, the Normal episode in 652-897AD and the Dry episode in 898-1320AD are similar among the three series (Fig. 4.8). The Wet episode in 1320-1490AD and the Dry episode in 1491-1709AD in the precipitation reconstruction (Fig. 4.8c) are roughly comparable to those in the $\delta^{18}O$ series (Fig. 4.8b). The Severely Wet episode in 1710-1840AD in the precipitation reconstruction (Fig. 4.8c) is similar to that in the TIC series (Fig. 4.8a).

Interestingly, the period from 1479 to 794BC, in which the TIC and $\delta^{18}O$ series were disconnected between the piston cores PLC98-4 and PLC 97-1, is identified as a period of very strong changes on short time-scales in the precipitation reconstruction (Fig. 4.8c).

### 4.5 VERIFICATION WITH RELATED STUDIES

Although the precipitation reconstruction is for the Climate Division 3 in Nevada, the tree-ring samples of composing the unfiltered precipitation reconstruction were collected from the White Mountains in California. Moreover, because the significant changes identified in this section are presented on multicentennial timescales, climate changes
usually occurred in a large geographical area, especially during dry periods. Thus we can consider the changes in the precipitation reconstruction series as an epitome of what happened in the western USA. Given this, we may try to use previously published archaeological reports about that region to further collaborate our partition results.

Some related papers published in the last 30 years have examined climatic fluctuations in precipitation or air-temperature in the western USA based on analyses of sediments, pollen, vegetation and tree rings besides Hughes & Graumlich [1996] and Benson et al [2002]. For example, Madsen et al. [2001] analyzed remains of small animals from stratified raptor deposits together with fossil woodrat midden samples to partition climate epochs in the eastern Great Basin during the late Pleistocene and Holocene. Grayson [2000] interpreted a decrease in small mammal fauna in the Bonneville Basin of north central Utah as evidence of a dry climate during the Middle Holocene. Benson et al. [1997] investigated Δ¹⁸O and TIC changes in Owens Lake of the Great Basin for the period from 17000BP to 4500BP. Feng & Epstein [1994] studied a hydrogen isotope time series covering the last 8000 years from Bristlecone Pines in the White Mountains of California. Stine [1994] synthesized data from relict tree stumps in Mono Lake and Tenaya Lake in California, and in southernmost Patagonia in South America (48° S - 50° S) to determine extreme and persistent drought in California and Patagonia during medieval times. Stine [1990] presented lake-level fluctuations in Mono Lake, California, for the last 4000 years. Wigand [1987] described the changes in vegetation for the last 6200 years at Diamond Pond in the eastern Oregon desert. However, no report of variance changes was found in these researches, so we discuss only changes of episode means in this section.

For comparison to these studies, we take 1950 as the present year to convert years before present (BP) into calendar years (BC/AD). As mentioned above, sediment cores and pollen records have a dating accuracy of about 50-100 years even more in multi-millennial chronology. The records from cores also may be biased in reflecting the extreme events of climatic droughts or floods, rather than the average conditions over multicentennial periods. These will make the alignment of dates only approximate and for rough, but useful, verifications of the climate situations.

The middle Holocene period from 6000BC to 3300BC is commonly recognized to be dry and warm in the Great Basin area, which covers most of Nevada and neighboring regions of southeastern Oregon, eastern and southeastern California, and western Utah [Benson, et al. 2001; Madsen, et al. 2001; Grayson, 2000; Feng & Epstein, 1994; Wigand, 1987]. Five of the eight extreme droughts were distinguished in around 5970, 5881, 5591, 4058 and 3948BC respectively by Hughes & Graumlich [1996]. This epoch coincides also the global first long warm phase during the last ten thousand years, known as the Europe Climatic Optimum period [Asakura, 1991]. In our analysis (Table 4.1), the Severely Dry episode in 6000-5866BC and Dry episodes in 5706-5340BC, 4313-3771BC occurred in this period. Moreover, above-mentioned five of the eight years of extreme droughts in Hughes & Graumlich [1996] are included in these three Dry episodes.

In China, the climate in this period also featured warm and wet, glaciers retreated in the western mountains while the desert shrunk in Inner-Mongolia. The Lake Daihai (112°39'E, 40°30'N), for example, was four times the area of that at present, and the "Painted Pottery Literature" developed in the middle-reaches of Yellow River in China during this period [Ye
However, there exist comparatively short wet spells during this long warm phase. For instance, most glaciers on the Earth progressed around 5400BC [Goodess, et al. 1992]. Around 5700BC the Great Salt Lake level had ever risen up to its normal level about 1283m, which was estimated from marsh deposits [Madsen et al., 2001; Murchison & Mulvey, 2000].

This may confirm the Wet episode in 5865-5707BC in our Table 4.1. There is a good fit to the Normal episode in 5339-4607BC (Table 4.1) by Madsen et al [2001], who quoted that between 5450BC and 4750BC [Rhode & Madsen, 1998] single-leaf pinyon nut hulls first appeared in the archaeological record from the Danger Cave (see Fig.4.1) in the west of the Great Salt Lake Desert, it suggests a wetter climatic environment than earlier. Other reports by Madsen et al. [2001] and Madsen [1985] state that a spell of greater effective moisture is indicated by a pronounced increase in the abundance of pine at the Potato Canyon Bog (see Fig.4.1) in central Nevada between 4550BC and 4050BC, this agrees with the Wet episode in 4606-4314BC (Table 4.1).

In China, two comparatively low magnetization-rates were recorded around 5330BC and 4700BC in the loess sediments at Baxie (103°24'E, 36°42'N) in Gansu province, it denotes relatively cold and arid climate [Ye & Chen, 1992]. Fang et al. [2004] summarized cold events in around 5450BC, 4750BC and a cold period from 4450BC to 4250BC in China based on a statistical analysis of cold events or period based on 97 investigation articles published before.

Wigand [1987] classified the phase from 3510 to 1850BC into the first wet period that heralded the end of the mid-Holocene drought, it is evidenced that sagebrush pollen increases in perennial Diamond Pond (Fig. 4.1) in the eastern Oregon desert, and that the littoral and aquatic plant macrofossils and mollusk shells appeared with sudden abundance shortly before 3510BC at Malheur Maar (Fig.4.1) in the east of the Oregon. Madsen et al. [2001] stated that between 3350BC and 2450BC, there was an increase in artiodactyl fecal pellets at Homestead Cave (Fig.4.1) in Utah, and quoted a mean lake-level elevation of 1280m or lower in the Great Salt Lake [Murchison & Mulvey, 2000] and markedly cooler conditions after 3350BC at Snowbird Bog (Fig.4.1) in Utah [Madsen & Currey, 1979]. These suggest a normal or wetter climate, and lend some credence to the Normal episode in 3770-3206BC and the Wet episode in 3205-2646BC in our Table 4.1. Moreover, most glaciers on the Earth developed again between 3500BC and 2400BC [Goodess, et al. 1992]. Fang et al. [2004] identified a cold period from 4050 to 3450BC and a cold event in 2950BC in China.

For the Dry episode in 2645-2383BC in our Table 4.1, Figure 24 in Wigand [1987] shows decreases in juniper pollen percentage and in the ratio of grass to sagebrush pollen, which mark a drier climate during this spell though accompanying description lacked in the text. Benson et al. [2002] summed up findings by Long & Rippeteau [1974], Hattori [1982], Wigand & Mehringer [1985], Grayson [1993], Stine [1990], and others, that there was a wet phase from 2300BC-1700BC. This is consistent with the Wet episode in 2382BC-1883BC in Table 4.1. Madsen et al. [2001] sorted the period from 2450BC to 1000BC as cooler and higher water levels in lakes, of which the Great Salt Lake level peaked up to 1284m in 1450BC [Murchison, 1989], and the following period from 1000BC to 450BC was regarded as much wetter and cooler. Wigand [1987] found that the period from 2050BC to 450BC was very wet with the deepest late-Holocene pond (Diamond pond) in around 1750BC. These descriptions
have some correspondence with the Normal episode in 1882BC-1418BC, the Wet episodes in 1417BC-1275BC and 1078BC-785BC in our Table 4.1, of which the Wet episode in 1078BC-785BC also coincided roughly with glacier advances in the Northern Hemisphere [Goodess, et al. 1992]. Bond et al. [1997] found two cold events in around 2350BC and 850BC in the North Atlantic by analyzing concentration of lithic grains and petrologic tracers in Holocene sediments of two cores from opposite sides of the North Atlantic.

In China, there were cold periods from 2050BC to 1750BC, from 1450BC to 1250BC and from 950BC to 750BC [Fang et al. 2004], in which the Hanshui River had frozen twice in 903BC and 897BC respectively [Zhu, 1979].

The Severely Dry episode 1274BC-1079BC (Table 4.1) contains the year 1251BC, one of the eight extreme dry years found by Hughes & Graumlich [1996]. No report of cold events was presented for the years between 1200BC and 1000BC either by Bond et al. [1997] or by Fang et al. [2004].

It was known as an optimal period of a warm and wet environment in China [Zhu, 1979], and was found that there were elephants in the north of Henan province in south of Yellow River in eastern China during the spell from 1300BC to 1100BC [Ye & Chen, 1992].

Though the Severely Dry episode in 784BC-279BC in Table 4.1 diverges from the generally wet conditions for this period suggested by Madsen et al. [2001] and Wigand [1987], Wigand [1987] noted a brief but significantly drier period after 650BC, in which it was reflected by less-abundant floating and submerged aquatic plants at Diamond Pond. During this time most glaciers on the Earth receded [Goodess, et al. 1992].

In China, the climate returned to a warm phase from 700BC to 20AD [Zhu, 1979], but with a relative cold period from 350 to 250BC [Fang et al. 2004].

The Dry episodes of 458AD-651AD and 1491AD-1709AD in Table 4.1 are also recognizable as dry periods by Hughes & Graumlich [1996] as well as by Hughes & Funkhouser [1998], who showed that there was a greater incidence of intense persistent moisture deficits after 400AD and before 1500AD in the Great Basin of North America. However, a cold event in 550AD in the North Atlantic was reported by Bond et al. [1997].

For the Normal episode in 652AD-897AD in Table 4.1, Madsen et al. [2001] give evidence that around 750AD a kind of fish, Utah chub, thrived and that hackberry endocarps were common, which indicate significantly moister conditions in the Homestead Cave vicinity. During this period most glaciers advanced again on the Earth, and the summer in Europe and America was comparatively cold [Goodess, et al. 1992].

In China, however, it was relatively warm between 600AD and 1000AD, droughts occurred in the middle-reaches of Yellow River and no snow and ice were seen in Xi’an city, the Capital of Shanxi province, for several winters [Zhu, 1979, Ye & Chen, 1992], but Fang et al [2004] concluded a cold event in around 850AD.

Previous investigations also appear to be consistent with the Dry episode in 898AD-1319AD in Table 4.1. For example, Stine [1994] designates the years from 900AD to 1200AD as a period of extreme and persistent drought in California, based on analysis of the tree-rings at Mono Lake (Fig.4.1). Madsen et al. [2001] summarized that between 1250AD and 1320AD, widespread droughts caused people to shift to full-time foraging in the Bonneville Basin [Madsen & Simms, 1998], and the period from 950AD to 1320AD may have been one of the warmest and driest phases in the Holocene [Harper & Alder 1970, 1972]. Wigand [1987] concluded that around 1250AD and 1450AD there were two major
droughts indicated by increases of greasewood values in Diamond Pond sediments. Wigand’s drought in 1450AD is just earlier than the Dry episode in 1491AD-1709AD in our Table 4.1 by less than 50 years, which might be considered to be coincided each other. The period from 900AD to 1300AD is called the Medieval Warm Epoch in the global change literature, the second warm phase during the last ten thousand years [Asakura, 1991]. Most glaciers on the Earth retreated during this phase [Goodess, et al. 1992].

In China, however, the average temperature was a little lower than at the present, though undergoing shorter fluctuations from warm (600AD to 1000AD), to cold (1000AD to 1200AD), to warm again (1200AD to 1300AD), and to another cold (1301AD to 1600AD) [Zhu, 1979], so that Fang et al [2004] classified the period from 1150AD to 1850AD as a cold phase.

There is collaborative evidence for the Severely Wet episode in 1710AD-1840AD in Table 4.1. Wigand [1987] discerned 1650AD–1800AD as wet with abundant juniper and grass pollen. Feng & Epstein [1994] concluded that there was a cool climate spell peaking between 1700AD and 1900AD. This period corresponds to the Little Ice Age in Europe, and the glaciers advanced on the Earth [Goodess, et al. 1992]. Bond et al [1997] also reported a cold event in around 1650AD in the North Atlantic.

China also experienced its coldest phase from 1601AD to 1899AD during the last 5000 years [Zhu, 1979, Ye & Chen, 1992].

It is easy to understand that the last episode since 1841AD is in the Normal category (Table 4.1), because the precipitation reconstruction is based on a regression analysis between the tree-ring records and the precipitation observations in the Nevada Climate Division 3 during the same period from 1932AD to 1979AD [Hughes & Graumlich, 1996]. This episode roughly corresponds to the third warm phase after 1850AD in the global change during the last ten thousand years [Asakura, 1991]. The declining trend (increase dryness) in the low-pass filtered curve of the precipitation reconstruction (Fig. 4.8c) for the last 50 years might reflect the global warming in the last century.

The warming in China for the last 50 years is appeared obviously in the winters and in the Northern China [Qing, et al. 2005].

Above-mentioned collaborating evidence is listed in the fifth column “quoted references” of Table 4.1. Only the Severely Wet episode in 278BC-457AD is not collaborated by related publications, but it is close to similar changes appeared in the $\delta^{18}O$ and TIC series. Also the glaciers mostly advanced on Earth in this period [Goodess, et al. 1992].

China transformed into the second cold phase (20AD to 600AD) during the last 5000 years [Zhu, 1979], and Fang et al. [2004] reported a cold period from 150AD to 550AD in China.

4.6 SUMMARY

Both algorithms, the scanning t-test and the scanning F-test, were applied to the 8000-year series of annual precipitation reconstruction from tree-rings in the southwestern USA. The precipitation reconstruction is calibrated in regression of tree-ring chronology to annual (prior July through current June) precipitation observations in region of the Climate Division 3 in Nevada. Based on the scanning t-test, twenty two significant change-points were identified in the subseries means and 23 wetness episodes were partitioned in the
precipitation reconstruction series. All episodes were classified into 5 grades according to the episode mean values of the precipitation reconstruction and are characterized in steady or unsteady features by combining with the results from the scanning $F$-test, which detects significant changes of subseries variances or standard deviation.

The coherency detection of significant changes in subseries mean was employed to compare the episodes partitioned in the precipitation reconstruction series with those in the TIC and $\delta^{18}O$ records, which were derived from Pyramid Lake sediment cores in the northwest of the Nevada. The algorithm was modified to accommodate unequal time intervals in the TIC and $\delta^{18}O$ time-series. It is shown that 13 of the 23 wetness-episodes in the precipitation reconstruction are approximately coincident with those in the TIC and $\delta^{18}O$ records.

Collaborating evidence from related paleoclimate studies was found for 22 of the 23 episodes. The related paleoclimate studies involve analyses of tree rings, pollen content, vegetation history, animal remains, and the TIC and $\delta^{18}O$ records in sediment cores collected from Nevada and vicinity states. All wetness-episodes, which were partitioned in the precipitation reconstruction series, are reasonably confirmed either by coherency of significant changes to those in the TIC and $\delta^{18}O$ records in Pyramid Lake sediments or by related previously published investigations (Table 4.1).

There seems to be a good relationship of the wetness-episodes in the precipitation reconstruction series with the periods in the global change: the warm and dry episodes in the southwestern USA usually coincide with the global warm phases, while the cold and wet episodes in the southwestern USA often associate with the global cold epochs. Scientists interpreted the global warm phases before 1850AD were controlled by the geomagnetic effect and changes in solar activities as well as the interactions between the atmosphere and ocean – ice – land, while the last warming period after 1850AD, especially for the last 50 years, is affected by human activities – mostly fossil fuel combustion [Eddy, 1976, Hood & Jirikowic, 1990, Goodess, et al. 1992, IDAG, 2005].

Persistent droughts and pluvials over the Plains and west of USA were recognized being ultimately driven by the tropical Pacific SST variations [Seager, et al. 2005]. The IDAG (International Ad Hoc Detection and Attribution Group, Zwiers, F, et al, 2005) summarized that the combination of La Niña and reduced moisture supply from the Gulf of Mexico likely led to the severe North American drought. Ropelewski & Halpert [1986] and Hidalgo & Dracup [2003] concluded that in general, southwestern U.S. cold season precipitation tends to be wetter than normal during El Niño events (negative phase of the Southern Oscillation), while drier than normal during La Niña events (positive phase of the Southern Oscillation). But the opposite effect is observed for the northwestern USA, creating a bipolar response between the two regimes.

Climate changes in China, are similar to the global change in general, but a little complicated in the some detail episodes such as a few spells during the period from 600AD to 1600AD. The effects of the El Niño and La Niña events on climate variation in China are also more complex than that in the southwestern USA [Qing, et al., 2005].

This work verifies that the tree ring series provide a valuable record of precipitation changes on climatic-scales in the southwestern USA, except perhaps for multi-millennial trends, which are more obvious in the TIC and $\delta^{18}O$ records. It also suggests that the algorithms of the scanning $t$-test, the scanning $F$-test and the coherency detection produce objective detection of multi-scale significant changes in both of subseries means and
subseries variances in a long time series, and of in-phase or out-phase significant changes between two time series, even when they are sampled on unequal time intervals.

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