

Temporal stability of climate-isotope relationships in tree rings of oak and pine (Ticino, Switzerland)

C. E. Reynolds-Henne,¹ R. T. W. Siegwolf,¹ K. S. Treydte,² J. Esper,² S. Henne,³ and M. Saurer¹

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[1] Climate reconstructions based on stable isotopes in tree rings commonly rely on the assumption that climate-isotope relations are stable over time. However, studies of tree ring growth have revealed trends thought to result from either physiological changes or changes in the climate-growth relationship. We investigated whether or not similar trends exist for tree ring stable isotopic ratios using a statistical approach. Correlations between climate (temperature and precipitation amount) and tree ring cellulose $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ of oak and pine from Ticino, Switzerland, were calculated for the period AD 1660–2000. Climate calibration of tree rings was enabled by long-term monthly resolved temperature and precipitation data sets on the basis of instrumental and documentary proxy data. Overall, five findings have been identified: (1) Isotopic ratios in tree rings most strongly reflect conditions of the current growing season, (2) temporally stable climate signals are found in pine $\delta^{13}\text{C}$ only, (3) all other correlations between tree ring isotopes and climate are temporally unstable and characterized by shifts in correlation sign and strength, (4) climate signals in oak are strongest in the 20th century, and (5) tree ring $\delta^{13}\text{C}$ reflects local climatic conditions while $\delta^{18}\text{O}$ is influenced by large-scale synoptic circulation. The nonstationary relationships observed could reflect changes in the relationship between the climate variables or a physiological adaptation to warmer conditions. Our results provide a cautionary note for the calibration of long tree ring series with 20th century relationships, at least for trees located at ecologically nonextreme sites.

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1. Introduction

[2] Historically, climate has strongly influenced human activity. Only recently, however, has the impact of human activity on climate come into question, in particular through the effect of increasing greenhouse gases during the last century [IPCC, 2001, 2007]. So-called natural climate variability, e.g., caused by changes in solar irradiance and volcanic eruptions [Lean *et al.*, 1995; Crowley and Kim, 1996] is superimposed on anthropogenically induced climate forcing. Disentangling the role of human and natural sources of variability within the climate system is essential in order to predict and cope with future climate scenarios. This might be achieved best by placing current changes into a long-term, historical context with which to evaluate future

trends [e.g., Esper *et al.*, 2002, 2005; Luterbacher *et al.*, 2004; Moberg *et al.*, 2005; D'Arrigo *et al.*, 2006].

[3] Ideally, in order to answer the question of how climate has changed in the past, we require reliable measurements on a large spatial scale extending as far back in time as possible. In practice, since widespread instrumental data are only available for about the past century, we rely on proxy climate indicators to yield insights into long-term climate variations. Of the several existing natural climate archives (e.g., lake sediments and ice cores), tree rings have the advantage of being almost globally distributed and annually resolved for up to thousands of years [Fritts, 1976]. As crossroads for global carbon and hydrologic cycles, trees are sensitive to climate through factors affecting carbon fixation, water use, and water availability [Rozanski *et al.*, 1993]. In this framework, stable carbon ($\delta^{13}\text{C}$) and oxygen ($\delta^{18}\text{O}$) isotope ratios in tree ring cellulose provide a continuous and complementary record of environmental conditions during tree growth [e.g., Edwards and Fritz, 1986; Anderson *et al.*, 1998; Treydte *et al.*, 2006].

[4] The $\delta^{13}\text{C}$ of tree ring cellulose reflects factors affecting photosynthetic uptake of CO_2 into leaves, such as relative humidity, temperature, and atmospheric CO_2 con-

¹Paul Scherrer Institute, Villigen, Switzerland.

²Federal Research Institute WSL, Birmensdorf, Switzerland.

³Swiss Federal Institute for Materials Science and Technology (Empa), Dübendorf, Switzerland.

centration [Farquhar *et al.*, 1982]. Specifically, the $\delta^{13}\text{C}_{\text{plant}}$ in plant matter is determined by $\delta^{13}\text{C}_{\text{atm}}$ of atmospheric CO_2 , by means of the following equation:

$$\delta^{13}\text{C}_{\text{plant}} = \delta^{13}\text{C}_{\text{atm}} + a + (b - a) \times (c_i/c_a) \quad (1)$$

where a represents the isotopic fractionation due to diffusion of CO_2 through the stomatal aperture (-4.4%), b is the fractionation due to enzymatic preference for ^{12}C during carboxylation (-28%), c_i is the partial pressure of CO_2 inside the leaf, and c_a is the partial pressure of CO_2 in the atmosphere [Farquhar *et al.*, 1989]. Thus the process of plant-atmosphere exchange is modulated by stomatal conductance so that water use efficiency is reflected in tree ring $\delta^{13}\text{C}$. Other factors like nutrient status and light availability influence carboxylation rates and photosynthesis and may site-specifically alter this relationship. However, consistent positive correlations with temperature and negative correlations with water availability have been observed across temperate, arid, and high-alpine sites [Leavitt and Long, 1988; Robertson *et al.*, 1997; Saurer *et al.*, 1997b; McCarroll and Pawellek, 2001; Treydte *et al.*, 2001].

[5] The $\delta^{18}\text{O}$ of tree ring cellulose mainly depends on the isotopic composition of water used during cellulose synthesis ($\delta_{\text{source}} = \delta_s$). Patterns in $\delta^{18}\text{O}$ of meteoric precipitation reflect evaporation and condensation within air masses at varying latitudes and distances from ocean sources on both spatial and temporal scales [Dansgaard, 1964; Siegenthaler and Oeschger, 1980; Rozanski *et al.*, 1993; Cole *et al.*, 1999; Bowen and Wilkinson, 2002]. Locally, the $\delta^{18}\text{O}$ signal of tree ring cellulose integrates stomatal conductance, which is coupled with transpiration [Yakir *et al.*, 1990; Barbour *et al.*, 2004] and is strongly related to relative humidity of the atmosphere (rh). On the basis of equilibrium isotope fractionation (ε_e) between liquid water and vapor, kinetic fractionation during diffusion of vapor (ε_k) and biochemical fractionations during cellulose production (ε_c), and a dampening factor f describing the mixing of differently enriched oxygen pools in the trunk, the following simplified model describes the isotope composition of tree ring cellulose [Saurer *et al.*, 1997a]:

$$\delta^{18}\text{O}_{\text{tree ring}} = f[\delta^{18}\text{O}_s + \varepsilon_k + \varepsilon_e + (\delta^{18}\text{O}_{\text{vapor}} - \delta^{18}\text{O}_s - \varepsilon_k) \times \text{rh}] + \varepsilon_c \quad (2)$$

[6] Previous studies have shown positive correlations between $\delta^{18}\text{O}$ of tree rings and local temperature [Saurer *et al.*, 1997a; Anderson *et al.*, 1998] and negative correlations with relative humidity [Switsur and Waterhouse, 1998; Saurer *et al.*, 2002] and water availability during the growing season [Saurer *et al.*, 1997b, Treydte *et al.*, 2006]. Precipitation amount, the calibration moisture proxy, relates more to the latter variable and therefore should also be negatively correlated with $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$.

[7] Inherent to the use of tree ring isotopes as climate proxies is the assumption that the tree ring–isotope and climate relationship is stable and predictable over time, allowing past climate to be inferred from calibrations with

recent instrumental data. A similar premise exists for the tree ring growth and climate relationship (ring width and maximum density), referred to as the “uniformitarian principle” [Fritts, 1976]. However, several studies have challenged this idea by showing a change in tree response to climate forcing in recent decades by reduced sensitivity to temperature [e.g., Briffa *et al.*, 1998; Büntgen *et al.*, 2006] and increased sensitivity to summer moisture stress [Biondi, 2000]. On a longer timescale, Carrer and Urbinati [2004, 2006] found time-dependent growth sensitivities in alpine European larch: These increased for trees younger than 200 a and mainly remained constant for older trees. Explanations for such changes range from temperature-induced drought stress [Barber *et al.*, 2000] and potential (but debated) fertilization effect of CO_2 [La Marche *et al.*, 1984; Feng, 1998; Knapp *et al.*, 2001] to falling ozone concentrations in the stratosphere [Bartholomay *et al.*, 1997; Briffa *et al.*, 2004], increased SO_2 emissions [Wilson and Elling, 2004], or tree age effects [Carrer and Urbinati, 2004]. Concerning stable isotopes, apart from an age-related effect caused by exposure to soil-respired $^{13}\text{CO}_2$ and lower light levels in juvenile trees, affecting $\delta^{13}\text{C}$ values in innermost tree rings [Francey and Farquhar, 1982; Schleser and Jayasekera, 1985], no time-dependent changes in climate sensitivities should occur. However, significant differences in the temporal response of $\delta^{13}\text{C}$ indices to temperature in oak trees were observed for the 20th century [Aykroyd *et al.*, 2001]. If factors other than climate dominate isotopic forcing, then climate reconstructions based on 20th century calibrations will be poor. From a palaeoclimatic perspective this relationship needs to be investigated.

[8] This study assessed the stability of the relationship between tree ring cellulose $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ and climate (temperature and precipitation amount) over the period AD 1660–2000 for oak (*Quercus petraea*) and pine (*Pinus sylvestris*) growing south of the main crest of the Swiss Alps, in the canton Ticino. Chronologies were based on a pooled-ring approach with four trees. Several other studies support the pooling of rings from approximately four trees to develop an isotope time series representative of each site, free of genetic or tree-specific influences [Leavitt and Long, 1984; Borella *et al.*, 1998; McCarroll and Loader, 2004]. Our long-term analysis was possible because of the emergence of a new high-resolution long-term gridded climate data set for the Greater Alpine Region based on both instrumental data and documentary proxy evidence [Casty *et al.*, 2005]. The temporal change in climate-isotope relationships was assessed statistically by means of correlation functions for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ in tree rings of pine and oak and near-surface air temperatures and precipitation values with monthly resolution.

2. Materials and Methods

2.1. Sampling and Sample Preparation

[9] Sampling occurred within the EU project ISONET. The study area, located in Switzerland south of the Swiss Alpine crest, was selected in order to find 400-a-long tree ring series for both pine and oak species. Samples were collected at two locations: Vigera for pine (*Pinus sylvestris* L.) and

Table 1. Site Characteristics^a

Site	Species	Latitude, °N	Longitude, °E	Slope	Aspect	Altitude, m asl
Caverigno	<i>Quercus petraea</i>	46°21′	8°36′	40°	170°	900
Vigera	<i>Pinus sylvestris</i>	46°30′	8°46′	35°	200°	1400

^aThere are n = eight sampled cores (four trees) per site; asl: above sea level.

Caverigno for oak (*Quercus petraea* (Matt.) Liebl) (Table 1 and Figure 1).

[10] The main factor distinguishing the biophysical environment of both sites is altitude (Table 1). The site at Vigera is an open subalpine coniferous forest with a shallow, podzolic ranker soil (30–40 cm) and includes *Picea abies* L. karst as a codominant species. The site at Caverigno is an open forest stand growing on rocky rendzina soil (40–50 cm) and includes *Fagus sylvatica* L. as a codominant species. At the latter site, logging was prohibited after about AD 1600 to protect the town of Caverigno from rock slides and landslides. Therefore this forest is one of the oldest

oak stands in the Greater Alpine Region (G. Carrero, personal communication).

[11] At both sites, care was taken to select trees under similar growth conditions. Living trees were cored twice at breast height; four trees per site were selected for isotopic analysis, while 26 (pine) and 18 (oak) trees were used for additional ring width analysis. Only the oak trees were mature at the start of the study period in AD 1660: Series began in AD 1560, 1596, 1617, and 1636 for oak and in AD 1651, 1653, 1664, and 1675 for pine. The nearest meteorological station Piotta (46°31′ N/8°41′ E, 1007 m above sea level (asl)) recorded average yearly temperatures of 7.2°C and total yearly precipitation of 1412 mm for the period AD 1979–2006.

[12] Ring widths were measured using a semiautomated RinnTech system (Heidelberg, Germany) with a resolution of 0.01 mm and cross-dated using the program COFECHA [Holmes, 1983]. Tree rings were split year by year with a scalpel; because of narrow rings, earlywood and latewood were not separated. Cross-dating ensured correct ring dating. Except for the period before AD 1680 (oak) and AD 1750 (pine), samples were pooled and single isotopic measure-

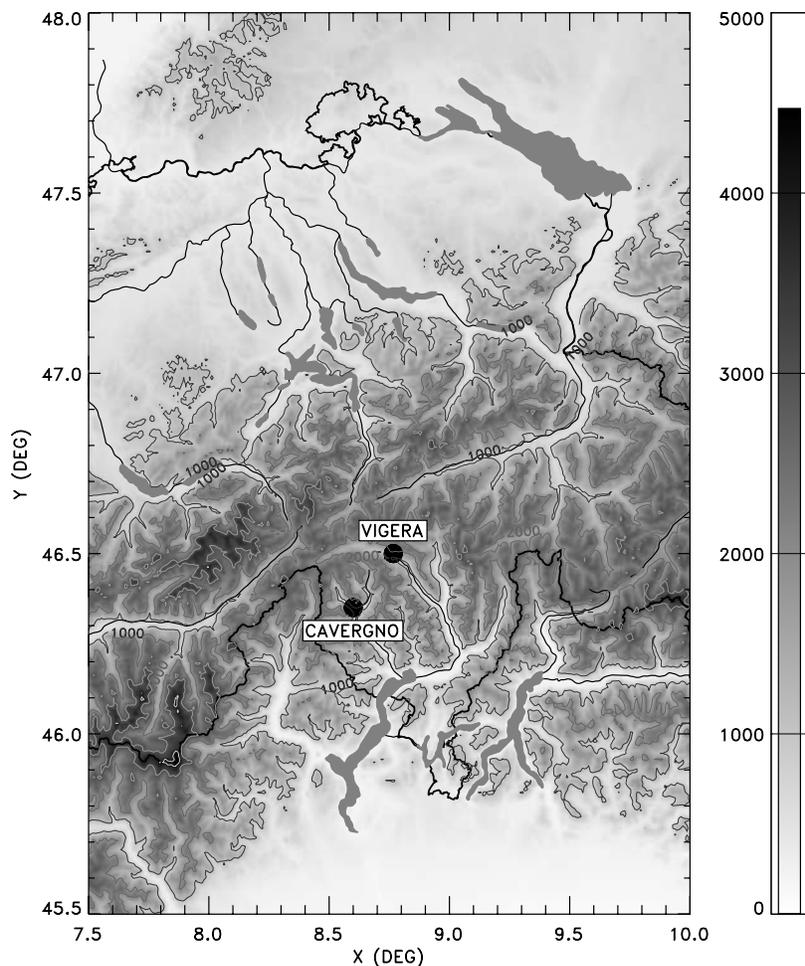


Figure 1. Map of sampling sites located south of the main crest of the Swiss Alps. Oak was sampled at Caverigno, and pine was sampled at Vigera. Reproduced by permission of the Swiss Federal Office of Topography.

ments per annum and species were conducted (two cores per tree and four trees, amounting to eight samples pooled for one analysis). The pooling approach was adopted for all the ISONET sites (23 chronologies) and proved successful for climate analysis [Treydte *et al.*, 2007]. Alpha-cellulose was extracted from milled samples following standard procedures [Loader *et al.*, 1997; Böttger *et al.*, 2007] and burned to CO₂ ($\delta^{13}\text{C}$) or pyrolyzed to CO ($\delta^{18}\text{O}$) prior to mass spectrometric analysis [McCarroll and Loader, 2004; Saurer and Siegwolf, 2004]. For carbon, isotopic ratios were determined by combustion in an elemental analyser (EA-1110, Carlo Erba Thermoquest, Milan, Italy) coupled to an isotope ratio mass spectrometer (Delta S, Thermo Finnigan Mat, Bremen, Germany). Reproducibility for the $\delta^{13}\text{C}$ analysis was 0.1‰. For oxygen, isotopic ratios were determined using a continuous flow method similar to carbon with the same mass spectrometer connected via a variable open-split interface ConFlo II (Thermo Finnigan Mat, Bremen, Germany). Reproducibility for the $\delta^{18}\text{O}$ analysis was 0.3‰. The isotope signature is expressed in the delta notation as

$$\delta^{18}\text{O} \text{ or } \delta^{13}\text{C} = (R_{\text{sample}}/R_{\text{standard}} - 1) \times 1000 \quad (3)$$

relative to an international standard (V-SMOW for oxygen, V-PDB for carbon) where R represents the ratio of heavy to light isotopes of a sample and standard, respectively. All carbon isotope series were corrected for the decrease in atmospheric $\delta^{13}\text{C}$ due to increased fossil fuel burning since the beginning of industrialization using ice core data [Leuenberger *et al.*, 1992; Francey *et al.*, 1999], complemented by measurements of atmospheric CO₂ samples for recent years [Allison *et al.* [2003], data from Mauna Loa]. A single isotope chronology was developed for the period AD 1660–2000 for each species.

2.2. Climate

[13] In general, climate in the Swiss Alps is influenced by interactions of synoptic weather systems originating from the North Atlantic, the Eurasian landmass, and the Mediterranean Sea [Wanner *et al.*, 1997]. For climate-isotope analysis, averaged monthly air temperature and total monthly precipitation sums for the geographic grid (46–46.5°N, 8.5–9°E) were obtained from the gridded data set “European Alps Temperature and Precipitation Reconstructions” with a 0.5° × 0.5° resolution [Casty *et al.*, 2005; Mitchell and Jones, 2005] for the period AD 1659–2000. The data set combines instrumental data extending back to AD 1676 (temperature) and AD 1688 (precipitation) with historical records, e.g., wine harvest dates, fluctuations of glacier tongues, long-lasting droughts, lake levels, and intense storms and floods [Brázdil *et al.* [2005], as discussed by Casty *et al.* [2005]], using principal component regression analysis. Historical records of both temperature and precipitation included in this reconstruction for the Swiss Plateau (~46.5°N, ~7.35°E, ~565 m asl) extend back to AD 1500. The oldest temperature measurements from Switzerland included in the Casty *et al.* [2005] data set are from stations in Geneva (AD 1752) and Basel (AD 1754), while the oldest precipitation measurements are from Zürich (AD 1708) and Bern (AD 1760). The earliest measured Swiss

Alpine temperatures included are from AD 1817 (Grand St. Bernard, 2472 m asl, 42.52°N, 7.1°E) and the earliest measured precipitation is from AD 1864 (Sils Maria i. E., 1798 m asl, 46.26°N, 9.46°E). The instrumental station number increases steadily from AD 1750 onward. In the 20th century, 98 predictors for temperature and 158 predictors for precipitation are used to construct this data set. Overall, this reconstruction indicates cold winters prior to AD 1900, followed by a warming trend until present day. While warmer summers are recorded for the second half of the 18th century, the warmest years since AD 1500 occurred in the 1990s. The reconstruction shows no significant low-frequency trend for precipitation. On a decadal scale, total growing season (MJJA) precipitation remains rather constant, though total wintertime precipitation (DJF) is relatively low from AD 1850–1900, relatively high around AD 1925, and decreases steadily from AD 1925 onward (Figure 2a). The long-term temperature course is characterized by growing season (MJJA) highs at ~AD 1800, 1870, 1950, and in recent decades, while winter temperatures (DJF) show a slow increase since AD 1775 interrupted by several cooling events but reaching higher levels today than any other period of the record (Figure 2b).

[14] From AD 1864 onward, daily temperature and precipitation data were available from the MeteoSwiss meteorological station Lugano (273 m asl, 46°14′N/8°57′37″E). It should be noted that these instrumental data are included in the Casty *et al.* [2005] data set. For the grid cell occupied by the trees studied, both sets of meteorological data (Lugano and Casty *et al.*) correlate well for the period AD 1864–2000 ($r = 0.91$ for temperature and $r = 0.77$ for precipitation, monthly values). The $\delta^{18}\text{O}$ isotopes in precipitation were available since 1973 from the GNIP station Locarno (46°10′N/08°48′E, Global Network of Isotopes in Precipitation [International Atomic Energy Agency (IAEA), 1992]; data from AD 1992–2003 provided by the Federal Office of Water and Geology (FOWG) and measured by the Climate and Environmental Physics Department of the Physics Institute of the University of Bern, Switzerland).

2.3. Climate-Isotope Analysis

[15] Annual isotopic data were calibrated against monthly temperature and precipitation data (AD 1864–2000, Lugano, MeteoSwiss station) using bootstrapped correlation analysis. Temporal changes in the relationship between climate and tree ring isotopes were analyzed using moving correlations in 40-a time windows on the basis of the Casty *et al.* [2005] data set, producing a time series of correlation coefficients. Correlations were calculated on a monthly basis from previous March (MAR) to current October (Oct), allowing seasonal sensitivity of correlations to be observed. In moving correlation functions the 40-a interval was progressively slid through time to compute correlation coefficients. Correlation coefficients were calculated for each monthly climate predictor (temperature or precipitation amount). Significance was tested using a bootstrap procedure [Guiot, 1991] in which 1000 random samples were drawn with replacement to obtain 1000 sets of correlation function estimates ($p < 0.05$). Mean correlation coefficients were considered significant if they were at least 2 times their

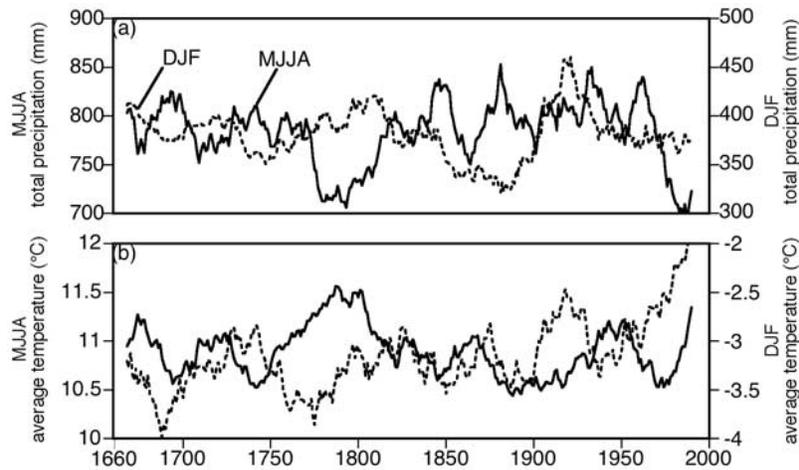


Figure 2. The 20-a running means of total precipitation (a) and average monthly temperatures (b) for May–August (MJJA, solid line) and December–February preceding the growing season (DJF, dotted line) at study sites in Ticino. Data are from *Casty et al.* [2005].

standard deviation. Statistical analyses were carried out using the computer program DendroClim [*Biondi and Waikul, 2004*] and R version 2.3, 2006.

3. Results

3.1. Raw Data

[16] First-order autocorrelations (lagged 1 a) are high for $\delta^{13}\text{C}$ series at both sites (0.69 and 0.55 for oak and pine, respectively), while lower autocorrelations for $\delta^{18}\text{O}$ (0.08 and 0.29 for oak and pine, respectively) were observed. A long-term increasing trend in oak $\delta^{13}\text{C}$ exceeding 1‰ is observed, and values for both species follow a similar

course after AD 1900 (Figure 3a). For $\delta^{18}\text{O}$, similar decadal-scale variations for both species are observed after AD 1750, with an offset of 2–3‰ (Figure 3b).

3.2. AD 1864–2003: Comparison With Instrumental Climate Data

[17] Results show that the amount of precipitation in current June and/or July determines the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ signals for this period (Figures 4a–4d). All four relationships between isotopes and precipitation show negative correlations for one or both of these months, while no other month is significantly correlated. Accordingly, there is no influence of the previous year. In contrast, for temperature,

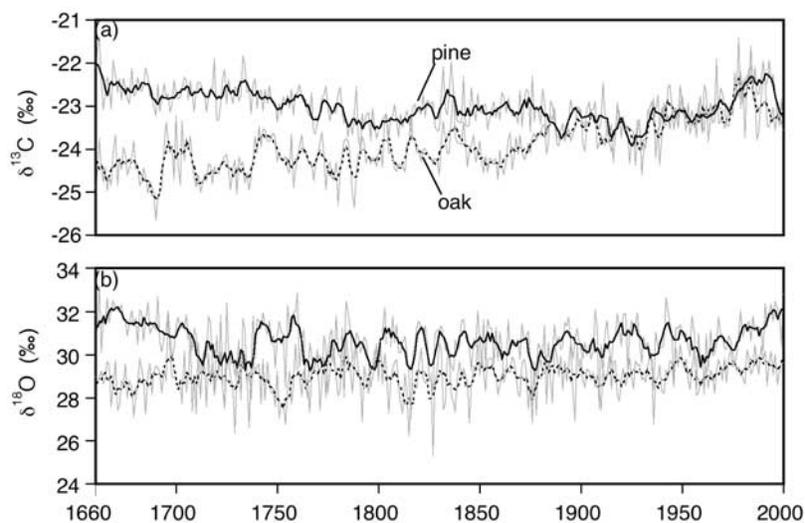


Figure 3. Single averaged chronologies for both pine (solid black line) and oak (dotted black line) for $\delta^{13}\text{C}$ (a) and $\delta^{18}\text{O}$ (b) for the period AD 1660–2000. Raw data are depicted in light gray, while black lines show 5-a running means of the same data. The $\delta^{13}\text{C}$ values were corrected for the trend in $\delta^{13}\text{C}$ of atmospheric CO_2 .

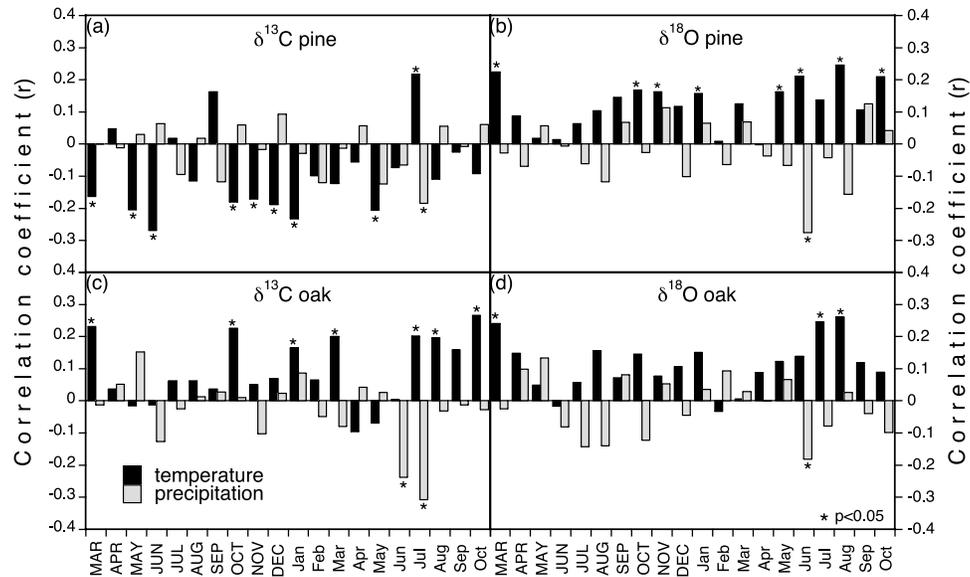


Figure 4. Bootstrapped correlation coefficients calculated for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of pine and oak and average monthly temperature (black bars) and precipitation amount (gray bars) of previous March (MAR) to current October (OCT) for the period AD 1864–2003. Climate data were obtained from the MeteoSwiss station Lugano.

several significant relationships ($p < 0.05$) are observed between monthly variations and isotopic variations. On average, significant correlations are found for current summer months and March of the previous year for both isotopes. Spanning the two growing seasons, the significant correlations with temperature are positive, with the exception of pine $\delta^{13}\text{C}$. To test the stability of the climate-isotope relationship over this period of high reliability of climatic data, we split the instrumental period into three subperiods (AD 1864–1900, 1901–1950, and 1951–2003) and performed correlation analyses. While no significant correlations between either temperature or precipitation and $\delta^{18}\text{O}$ (pine and oak) exist for the period AD 1864–1900 (MJJA), pine $\delta^{18}\text{O}$ shows strongest correlations with climate for the middle period (AD 1901–1950), whereas oak $\delta^{18}\text{O}$ shows strongest correlations with climate for the latest period (AD 1951–2003). For $\delta^{13}\text{C}$ the earliest period, AD 1864–1900, reveals overall strongest correlations between isotopes and both temperature ($r = 0.56$, pine, July temperature) and precipitation (e.g., $r = -0.46$, oak, June precipitation), while correlations are insignificant for all months for oak $\delta^{13}\text{C}$ and temperature. For pine $\delta^{13}\text{C}$ and precipitation, significant correlation coefficients decrease over time from the first to the last instrumental subperiod.

3.3. AD 1660–2000: Carbon Isotopes and Climate

[18] To investigate the temporal stability of the climate-isotope relationships over longer timescales, 40-a running correlations between the *Casty et al.* [2005] climate reconstructions and the isotope records were conducted (Figures 5a–5h), where each bar on the graph represents correlations calculated for different months of the growing season over a 40-a period, plotted at the last year of the period. Figures 5a–5h express significant correlation coef-

ficients on a color scale for studied months (previous March to current October; y axis) as a function of the investigated period (x axis). For pine $\delta^{13}\text{C}$, correlations with temperature are consistently positive for current growing season temperatures (mainly July and August) (Figure 5a). Prior to AD 1750, all but the months of previous August and current January are important in determining the $\delta^{13}\text{C}$ signal, while in the 20th century the current month of July strongly dominates.

[19] Correlations with precipitation amount are mostly negative (Figure 5b). For oak $\delta^{13}\text{C}$, correlations are more complex and show distinct temporal shifts (Figures 5c and 5d). Prior to AD 1825, correlations with temperature are equally significant and negative for both the current and previous springs (March and April). For the same time interval, significant correlations with precipitation are predominantly negative for the months of previous May to current February. A breakdown in this pattern occurs from AD 1825–1900, when “isotopically important” temperatures shift from springtime to current April–October. By AD 1900 the most consistent significant correlations with temperature are reduced to the months of current August–October, and the sign of the correlation changes from negative to positive. This coincides with a shift in correlations with precipitation amount to current year only (December–September) for the period AD 1900–2000. The temporal stability of the relationship between pine $\delta^{13}\text{C}$ and climate variables supports the reliability of the *Casty et al.* [2005] data set. If the *Casty et al.* [2005] data were unreliable in the early part of the series, then none of the isotope-climate relationships could be stable over the entire study period. Therefore the instability of the relationships between climate and the other isotopic chronologies

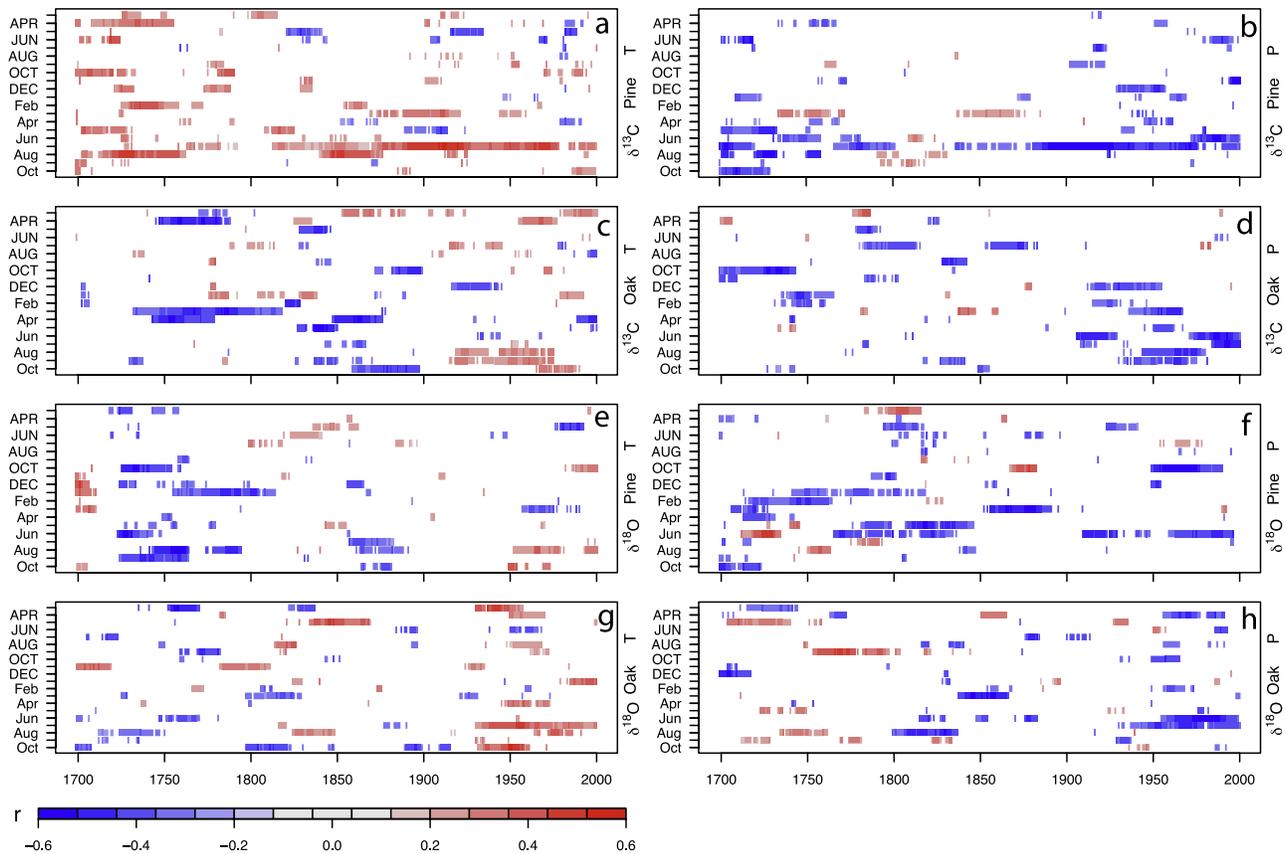


Figure 5. The 40-a running means of bootstrapped statistically significant correlation coefficients between climate data and tree ring isotopes for AD 1660–2000 (insignificant correlation coefficients set to “0”). Y axis: Correlations were calculated separately for each month for the period from March of the previous year (uppercase letters) to October for the current year (lowercase letters). Each bar represents a correlation calculated over a 40-a period plotted at the last year of each 40-a period. Color scale corresponds to both the sign and strength of the correlations.

observed are more likely related to changes in the climate-isotope relationship, rather than to uncertainties in the climate data set itself.

3.4. AD 1660–2000: Oxygen Isotopes and Climate

[20] For pine $\delta^{18}\text{O}$, despite significant correlations being spread over much of a given year, the current growing season dominates: June–September temperatures and May–June precipitation amounts mostly show significant correlations (negative before AD 1950) with $\delta^{18}\text{O}$ (Figures 5e and 5f). For the period prior to AD 1850, winter temperature and precipitation amounts are reflected in $\delta^{18}\text{O}$ signals more strongly than during the 20th century. In general, the number of significant relationships for both climate parameters decreases over time; for precipitation amount, after AD 1900 only 2 months stand out as important: current June and previous October (Figure 5f). For temperature, current growing season correlations are mostly negative up to AD 1900, followed by a shift to positive correlations for August and October for the 20th century. For oak the relationship between $\delta^{18}\text{O}$ and climate is highly variable prior to AD 1950 (Figures 5g and 5h). From AD 1950

onward both the strength and sign of the correlations become more consistent (positive correlations with July–October temperature, negative correlations with June–July precipitation amount). This coincides with the period of significant positive correlations observed between oak $\delta^{13}\text{C}$ and temperature (AD 1900 onward, Figure 5c) and between pine $\delta^{18}\text{O}$ and temperature (AD 1950 onward, Figure 5e). In addition, for the period AD 1973–2000, $\delta^{18}\text{O}$ in tree ring cellulose and $\delta^{18}\text{O}$ of precipitation (GNIP data) correlate during the growing season ($r = 0.6$ for both species).

3.5. Climate–Tree Ring–Isotope Relationships: Summary

[21] In Figure 6 the above results are summarized by showing the correlations to current and previous year growing season conditions for all eight relationships studied. Correlations for pine $\delta^{13}\text{C}$ and temperature (to a lesser extent precipitation) are temporally stable for the current growing season (Figures 6a and 6c). For oak $\delta^{13}\text{C}$, correlations are temporally unstable and weak (Figures 6b and 6d). Overall, the equal importance of the previous and current growing seasons in determining $\delta^{13}\text{C}$ signals dis-

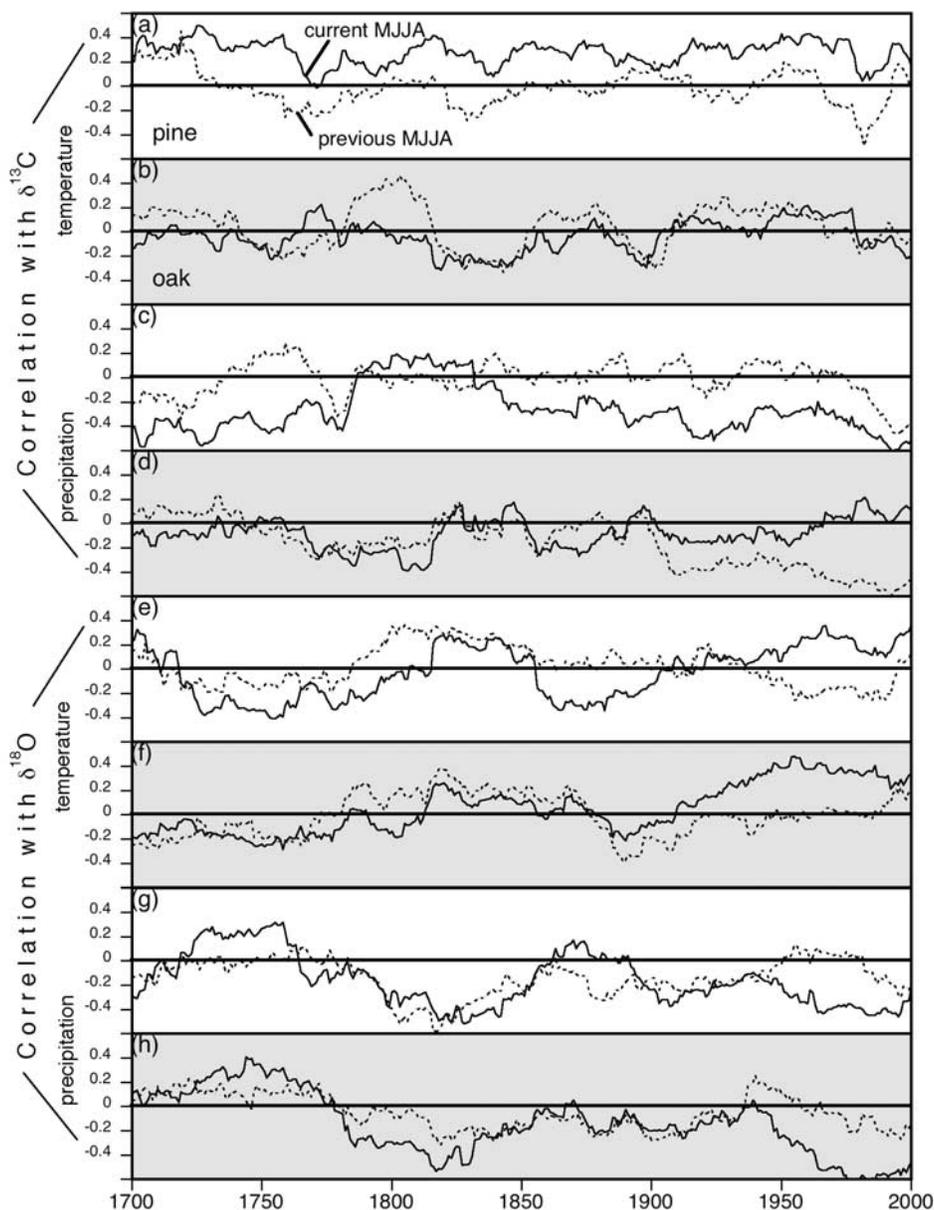


Figure 6. Correlation coefficient summary. The 40-a running means of bootstrapped correlation coefficients from AD 1660–2000 for average temperature (a, b, e, and f) and precipitation amount (c, d, g, and h) calculated for current MJJA (solid line) and previous MJJA (dotted line). Shaded graphs are results for oak, nonshaded are for pine. Note: This graph includes both significant and nonsignificant correlation coefficients.

tinguishes oak responses from pine. For $\delta^{18}\text{O}$, strikingly similar decadal and century-scale trends in relationships with both precipitation amount and temperature are observed during the growing season for both species (compare Figures 6e and 6f and Figures 6g and 6h). For instance, for precipitation, positive correlations with $\delta^{18}\text{O}$ are observed for pine and oak shortly after AD 1700, negative correlations are observed later on around AD 1800, increasing values are observed thereafter, and finally, strongly negative values are observed in the last 50 a. A similar pattern but

opposite in sign is observed for the correlations with temperature. Furthermore, similar to $\delta^{13}\text{C}$ the period after AD 1950 shows a change in the relative importance of previous and current growing seasons for oak for both climate parameters (loss of importance of previous May–August conditions over time).

4. Discussion

[22] The simple yet widely applied approach of correlation analysis for the instrumental period (AD 1864–2003)

yielded mostly expected results, namely, positive relationships with temperature for both isotopes and negative relationships with precipitation (Figure 4), despite correlations being relatively low overall. These relationships are expected on theoretical grounds. The positive relationship for $\delta^{18}\text{O}$ with temperature reflects the influence of source water ($\delta^{18}\text{O}$ of precipitation), while the negative impact of precipitation is due to the inverse relationship with air humidity (equation (2)). For $\delta^{13}\text{C}$ the physiological effect of drought, namely stomatal closure, is most expressed under the combined influence of high temperature and low precipitation, which results in the observed positive correlation with temperature and negative correlation with precipitation. The importance of the growing season was revealed by this analysis, although the influence of other months was also detected. Changing relationships for the subdivided period, however, produced some doubts concerning the stability of these relationships. Because of the relatively dense network of meteorological stations and available documentary proxy data in this region, it was possible to extend the correlation analysis over a longer period than is commonly used for calibration. This revealed that the results from the AD 1864–2003 analysis were not maintained throughout the longer period AD 1659–2000. Except for pine $\delta^{13}\text{C}$, relationships between climate and tree ring $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ are not stable over time but show significant shifts associated with the 20th century (e.g., changes in importance of previous and current growing season or changes in the sign of the correlation). These results suggest that changes in the climate forcing on tree ring isotopic composition could also be expressed through threshold-controlled mechanisms in trees, as for changes in growth sensitivity in the work of Carrer and Urbinati [2001, 2006]. While increased data resolution beginning in the 19th century in Switzerland could affect the strength of correlations, the instrumental age fails to explain the major changes in relationships observed from the start of the 20th century, as was also shown by poor correlations with isotopic data for the most recent part of the instrumental period (AD 1951–2003) following subdivision. Several possible explanations for the observed instability are discussed in the following paragraphs.

[23] Overall, the current growing season climate dominates $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ signals, despite significant relationships found in both species for several months from previous March to current October. For $\delta^{13}\text{C}$ the importance of both growing seasons is evidenced by high first-order autocorrelation functions. While current year climate is expected to be reflected in current year plant biomass including tree ring $\delta^{13}\text{C}$ (e.g., growing season precipitation amounts [Saurer et al., 1997b]), an impact on subsequent years may occur via the remobilization of stored sugars, production of buds and hormones, and growth of roots and root hairs [Fritts, 1976]. For oak, a species forming ring-porous wood, earlywood formation starts before bud burst and completes before leaf expansion, i.e., before new leaves become exporters of photosynthates [e.g., Pilcher, 1995]. Therefore this species must utilize stored carbohydrates to form new tissues resulting in potential climate signal carryover from 1 a to the next. Changes in the amount of

reserves used under different climatic regimes could potentially affect the isotope trends and shifting correlations. In contrast, coniferous species like pine produce almost the entire earlywood from current photosynthates [Dickmann and Kozlowski, 1970; Glerum, 1980]. In addition, for pines, while cold temperatures inhibit photosynthetic efficiency, they do not necessarily inhibit photosynthetic capacity, and carbon may be fixed year-round [Krol et al., 2002]. While the dominance of summertime precipitation is expected for $\delta^{18}\text{O}$ [Saurer et al., 1997b], there may be a depth-dependent temporal offset resulting from effective uptake of soil water by roots.

[24] AD 1900 marks several changes in the relationships between climate and $\delta^{13}\text{C}$. From this point onward a persistent increase in $\delta^{13}\text{C}$ occurs in both species over time, despite corrections accounting for atmospheric $\delta^{13}\text{C}$ decrease since industrialization (Figure 3). According to equation (1) this increase, observed in both species, is caused by a decrease in c_i/c_a , suggesting that CO_2 assimilation efficiency vis-à-vis water loss has increased over time, due to increased water use efficiency through CO_2 enrichment effects [Drake et al., 1997]. Several tree ring studies also concluded that the intrinsic water use efficiency of trees has increased over the last century [Waterhouse et al., 2004]. Concerning the relationships between climate and $\delta^{13}\text{C}$, a shift in the dominance of temperature from the month of August to July occurs, possibly related to a lengthening of the growing season [Keeling et al., 1996; Saxe et al., 2001]. Such a change in the seasonality of carbon uptake could induce a long-term shift in tree ring $\delta^{13}\text{C}$ due to the seasonality in isotopic composition of atmospheric CO_2 . Seasonal changes in atmospheric CO_2 $\delta^{13}\text{C}$ amounting to 0.2‰/month for European stations have been observed during summertime [Allison et al., 2003], which could contribute to changes in plant $\delta^{13}\text{C}$ according to equation (1).

[25] For oak a clear dependence on previous and current springtime temperatures at the beginning of the study period is replaced by positive correlations with current summer temperatures and a change in importance from previous to current year precipitation amounts. In contrast to $\delta^{13}\text{C}$, AD 1950 marks the change in relationships between climate and $\delta^{18}\text{O}$. First, the relationship between pine $\delta^{18}\text{O}$ and temperature changes from negative to positive for current summer–fall months. Second, the relationships between oak $\delta^{18}\text{O}$ and climate shift from random to consistent for both temperature and precipitation amount. While it is well known that the quality of instrumental records improves with time, the consistent and expected correlations obtained for pine $\delta^{13}\text{C}$ and climate provide evidence against the possibility that changes in the climate-isotope relationship over time could be due to uncertainties in the climate record [Casty et al., 2005].

[26] At dry sites, water conditions are more important than temperature in determining $\delta^{13}\text{C}$ in tree rings of beech [Saurer et al., 1995] and pine [Leavitt and Long, 1988; Gagen and McCarroll, 2004]. Steep slopes, shallow soils, and near-south exposure predestine the trees studied to water stress. However, the combination of cold and wet winters at high elevations causes an increase in the fraction of precipitation falling as snow, retarded snowmelt, and

increased water availability to roots during the growing season, as observed prior to AD 1800. Unlike the shallow root systems in pines, oaks generally form a taproot providing access to groundwater [Larcher, 1975] and are therefore slower to react to decreasing water availability as a result of warm winter temperatures (AD 1850 onward) and reduced winter precipitation (AD 1925 onward) (Figure 5). In addition, the lower elevation at the oak site suggests the stronger presence of nonclimatic factors affecting $\delta^{13}\text{C}$ such as competition or soil nutrient status, compared to pine. Coupled with age-related physiological changes in hydraulic status, these effects induce more stressful conditions in older trees [Carrer and Urbinati, 2004], thus explaining stronger correlations with climate in the 20th century.

[27] Furthermore, since the components of the climate system are all interrelated, changes in the relationship between the climate variables themselves over time (e.g., Figure 2) may affect correlation coefficient stability with isotopic series from trees. Edwards et al. [2000] have shown in controlled experiments with bean plants as well as in tree ring calibrations with *Abies alba* that the coupled influence of humidity and temperature may explain divergent correlations for carbon isotopes reported in the literature. Depending on the humidity regime, Edwards et al. observed very different slopes in the temperature-isotope relationship that are explained by independent δ -temperature and δ -relative humidity effects. When considering the long-term summer (MJJA) variations in precipitation (Figure 2a) and temperature (Figure 2b) in our region, one can observe a strong anticorrelation between the two climate parameters from the beginning of the record until about AD 1880 ($r = -0.79$), expressed particularly strongly at a high-temperature and low-precipitation anomaly around AD 1800. After AD 1880, however, this anticorrelation breaks down ($r = -0.11$), indicating a change in the relationship between the climate variables that may affect the isotope-climate relationships as well. In addition, nonlinear transformation of environmental signals through biological systems could be a reason for changes in temperature coefficients [Schleser et al., 1999].

[28] While averaged growing season trends reveal the nonstationary nature of the tree ring $\delta^{18}\text{O}$ -climate relationships (Figure 6), the highly similar pattern revealed by these smoothed curves at the two sites points to a common forcing factor. Correlations with $\delta^{18}\text{O}$ of precipitation (GNIP data) showed that source water signals are preserved in $\delta^{18}\text{O}$ of tree rings. A possible explanation for the common decadal-to-century-scale trends, not directly related to mean monthly air temperature or precipitation amount, is variations in moisture source determined by the dominant atmospheric circulation pattern [Dansgaard, 1964; Rozanski et al., 1993]. The North Atlantic Oscillation (NAO) index, defined as the normalized pressure differential between the Icelandic Low and the Azores High, determines dominant wind patterns and temperature and precipitation systems over Europe [Hurrell and van Loon, 1997] and influences $\delta^{18}\text{O}$ of precipitation in Switzerland [Teranes and McKenzie, 2001]. Winter (average DJF) NAO indices for the period AD 1659–2000 [Luterbacher et al., 2004] explain some of the variability in correlation coefficients calculated for

precipitation and $\delta^{18}\text{O}$ for both species ($r^2 = 0.17$ and 0.18 for oak and pine, respectively). Comparison of 40-a running bootstrapped correlations (isotopes versus climatic data, results from Figure 6) with 40-a running averages of $\delta^{18}\text{O}$ from a Greenland ice core (GISP2, Grootes and Stuiver [1997]) shows even more similarity ($r^2 = 0.52$ and 0.35 with correlation coefficients for oak and pine with precipitation, respectively). The mechanism behind this apparent teleconnection is not yet clear, in particular since we compared a derived variable to isotopic series from an ice core. Our results indicate that while oxygen isotopes remain a valuable proxy for precipitation as recently shown in High Asia [Treydte et al., 2006] and atmospheric circulation changes [e.g., Roig et al., 2006], modeling approaches covering larger scales might be needed to better decipher the climate signal [Hoffmann et al., 2000]. In contrast to $\delta^{18}\text{O}$, averaged growing season trends for $\delta^{13}\text{C}$ show no coherence between sites, reflecting the greater influence of local ecological conditions and physiological or age-related trends in determining $\delta^{13}\text{C}$ [Saurer et al., 1997a; Treydte et al., 2001, 2006].

5. Conclusions

[29] While long, continuous and absolutely dated isotope series from trees are considered a powerful tool for reconstructing past climate, the results of this study show that the assumption of temporal stability for climate–tree ring–isotope relationships is questionable. Long-term trends are not well represented by 20th century relationships, as sensitivity to climate conditions seems to increase strongly during this period of heightened anthropogenic influences. Therefore this period may not be the best with which to calibrate isotopic time series from trees. At alpine sites in Switzerland, $\delta^{13}\text{C}$ represents a more site-specific signal, while $\delta^{18}\text{O}$ shows larger-scale patterns of response to climate variability and change. The common nature of the $\delta^{18}\text{O}$ -climate response at two sites and the relationship with remote records of atmospheric circulation suggests the possibility for the reconstruction of synoptic weather patterns from tree rings and the potential for combining series of both oak and pine to extract a single climatic signal. Changing relationships between climate variables might be an important factor explaining instability in the climate-isotope relationship. Twentieth century shifts in the importance of previous and current year climate in determining $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ of oak imply that factors affecting isotopic signatures in trees strongly influence formation and use of stored compounds or the relative importance of earlywood and latewood during this period. We conclude that biophysiological growth may change to climate-limited growth in a very short period of time, even in mature temperate trees located at what are considered non-limiting sites. The implication for palaeoclimatic studies of tree rings is that these may be of varying reliability.

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J. Esper and K. S. Treydte, Federal Research Institute WSL, 8903 Birmensdorf, Switzerland.

S. Henne, Swiss Federal Institute for Materials Science and Technology (Empa), 8600 Dübendorf, Switzerland.

C. E. Reynolds-Henne, M. Saurer, and R. T. W. Siegwolf, Paul Scherrer Institute, 5232 Villigen PSI, Switzerland. (matthias.saurer@psi.ch)