



Summer relative humidity in northern Japan inferred from $\delta^{18}\text{O}$ values of the tree ring in (1776–2002 A.D.): Influence of the paleoclimate indices of atmospheric circulation

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[1] The summer relative humidity (RH) changes in Hokkaido, northern Japan, since 1776 were reconstructed using the oxygen isotope ratios of the tree ring cellulose of two living oak trees. We investigated the direct relationships between the decadal-centennial variations in the summer RH in northern Japan and the climate indices of atmospheric circulation to understand the factors affecting the changes in the hydrological climate in northern Japan. The variations in the summer RH are negatively correlated with those in the annual PDO indices since 1781. This is probably because the humid southerly wind from the western Pacific Ocean blows toward northern Japan with the intensified Pacific high when the PDO index is lower. Further, the fluctuations in the summer RH are positively correlated with those in the summer AO index during 1781–1930, but they are negatively correlated with those in the summer AO during 1940–1997. During the 1930s, the AO index changed from the negative to positive on the average. The drastic shift in its correlation is explained by the difference between atmospheric circulations in the low-AO period (1899–1930) and the high-AO period (1970–2000). The summer RH in northern Japan was regulated by the summer AO during 1781–1930 (the cold period) and the annual PDO during 1940–1997 (the warm period). As a consequence of global warming, the midlatitude forcing such as PDO might become stronger than the high-latitude forcing such as AO on the hydrological climate in northern Japan.

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1. Introduction

[2] More than half of the world's population lives in the Asian monsoon area, where huge floods often cause economic and agricultural losses and endanger the lives of people. In such areas, because of global warming, the future changes in the hydrological climate turn out to be one of the most crucial concerns. Climate models can only forecast the hydrological climate in the future. However, the observed hydrological climate data for approximately 100 years is insufficient to verify the climatic models.

[3] Data of the past long-term hydroclimate is needed and can be reconstructed by a climate proxy. There are some proxies of past changes in the hydrological climate, such as tree rings, ice cores, pollen analysis in the peat, and lacustrine sediments. Among the above proxies, the tree rings can hold high-resolution spatial and temporal records;

hence, they are appropriate for the purpose of extending the observed hydrological climate data back into the past. In Asia, the past hydrological climate changes have been reconstructed by the tree ring widths [Singh and Yadav, 2005; Pederson *et al.*, 2001; Bräuning and Mantwill, 2004; Sheppard *et al.*, 2004]. These studies were limited to the dry inner continental area, which is at the marginal region of the Asian monsoon area. This is because the tree ring width under the extremely arid climate is simply controlled by hydrological factors and therefore is useful for reconstructing the past changes in the hydrometeorological conditions. However, most of the Asian monsoon area belongs to the pluvial climatic division. In the dense forests in the pluvial monsoon areas, tree ring width is controlled not only by climatic factors such as temperature and precipitation but also by the physiological and ecological factors of the tree, such as the competitions for light in the environment. Consequently, it is difficult to use the tree ring width as the proxy for hydroclimate in the pluvial monsoon area. In the Asian pluvial monsoon area, including Japan, past long-term reconstructed records of the hydroclimate such as precipitation and relative humidity (RH) based on tree ring data are very limited.

[4] For over 30 years, the relationships between the oxygen isotope ratios of tree ring cellulose and climatic

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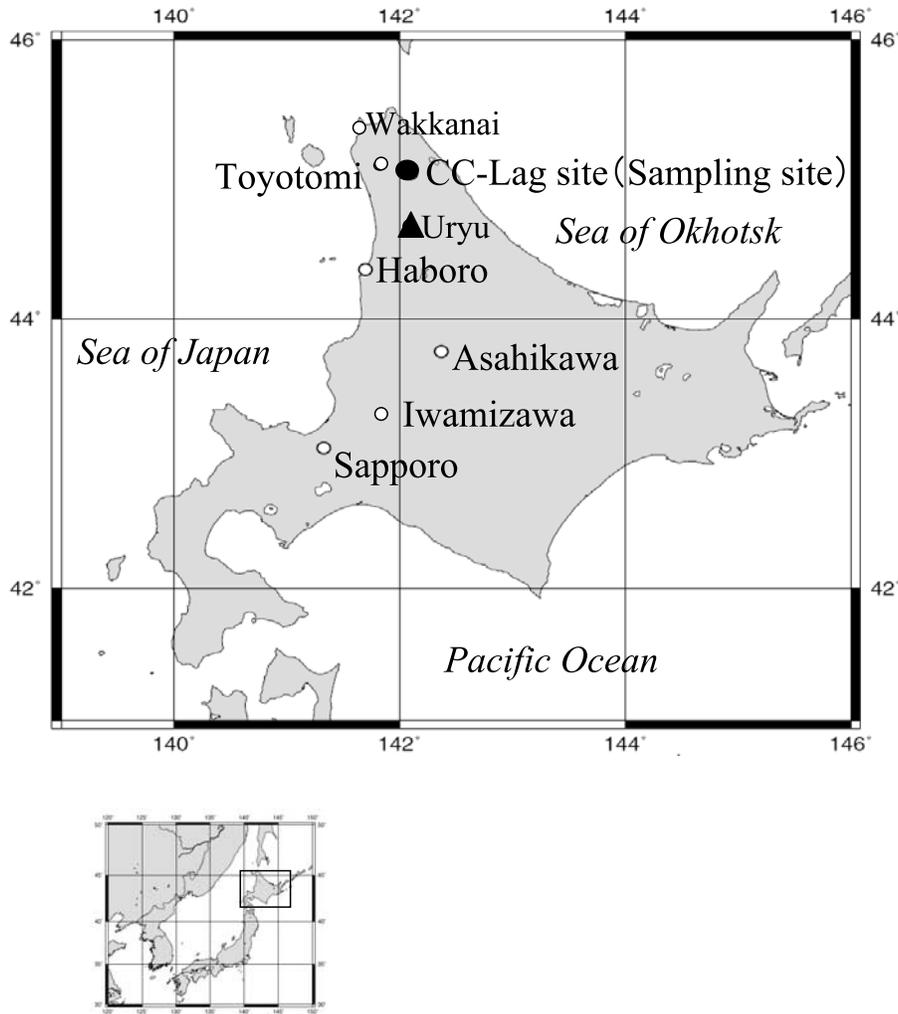


Figure 1. Map of Hokkaido, the northern part of Japan. Open circles, meteorological observatory (Sapporo, Asahikawa, Haboro, Iwamizawa, and Wakkanai); solid circle, CC-Lag site (sampling site); solid triangle, Uryu (sampling site of *Nakatsuka et al.* [2004]).

factors have been investigated in Europe and North America in order to use them to reconstruct climate [e.g., *Gray and Thompson, 1976*]. Recently, the precipitation over the last millennium in northern Pakistan, which is out of the monsoon area, was reconstructed by the oxygen isotope ratios of tree ring cellulose [*Treyde et al., 2006*]. In Asian pluvial monsoon areas, studies in the oxygen isotope ratios in tree ring cellulose are very scarce [*Nakatsuka et al., 2004; Tsuji et al., 2006*]. The oxygen isotope ratios of tree ring cellulose are mainly controlled by the oxygen isotope ratios of soil water and leaf water [*Roden et al., 2000*]. The oxygen isotope ratios of the precipitation during the growth season, which is taken up by trees as soil water, are negatively correlated with the monsoon precipitation amount in monsoon pluvial areas such as Japan, and positively correlated with the temperatures in the high latitudes [*Dansgaard, 1964*]. The oxygen isotopes of the leaf water is largely (poorly) enriched by transpiration with lower (higher) RH [*Dongmann et al., 1974*]. The oxygen isotope ratios of tree ring cellulose are controlled by RH and/or precipitation amount in the monsoon pluvial areas and the temperature in the high latitudes, respectively. The

mechanism determining the oxygen isotope ratios of tree ring cellulose is basically different from that used for determining the tree ring width.

[5] Since both precipitation amount and RH are negatively correlated with the oxygen isotope ratios of tree ring cellulose in the pluvial monsoon areas, these ratios of tree rings can be used as pure proxies of the dry–wet conditions at that location. Indeed, in Hokkaido, northern Japan, which belongs to the Asian pluvial monsoon area, it has already been shown that the oxygen isotope ratios of the tree ring cellulose of oak trees (*Quercus crispula*) are negatively correlated with the summer RH and precipitation [*Nakatsuka et al., 2004; Tsuji et al., 2006*]. Thus, we used the oxygen isotope ratios in the tree ring cellulose of two living, oak trees to reconstruct the variation in the past hydroclimate in Hokkaido, northern Japan, from since 1776.

[6] In addition we analyzed the reconstructed variation in the past hydroclimate in northern Japan by periodic analysis. We extracted the major component of variations in the reconstructed RH and investigated the relationships between the extracted major component in the reconstructed RH in northern Japan and the previously reconstructed

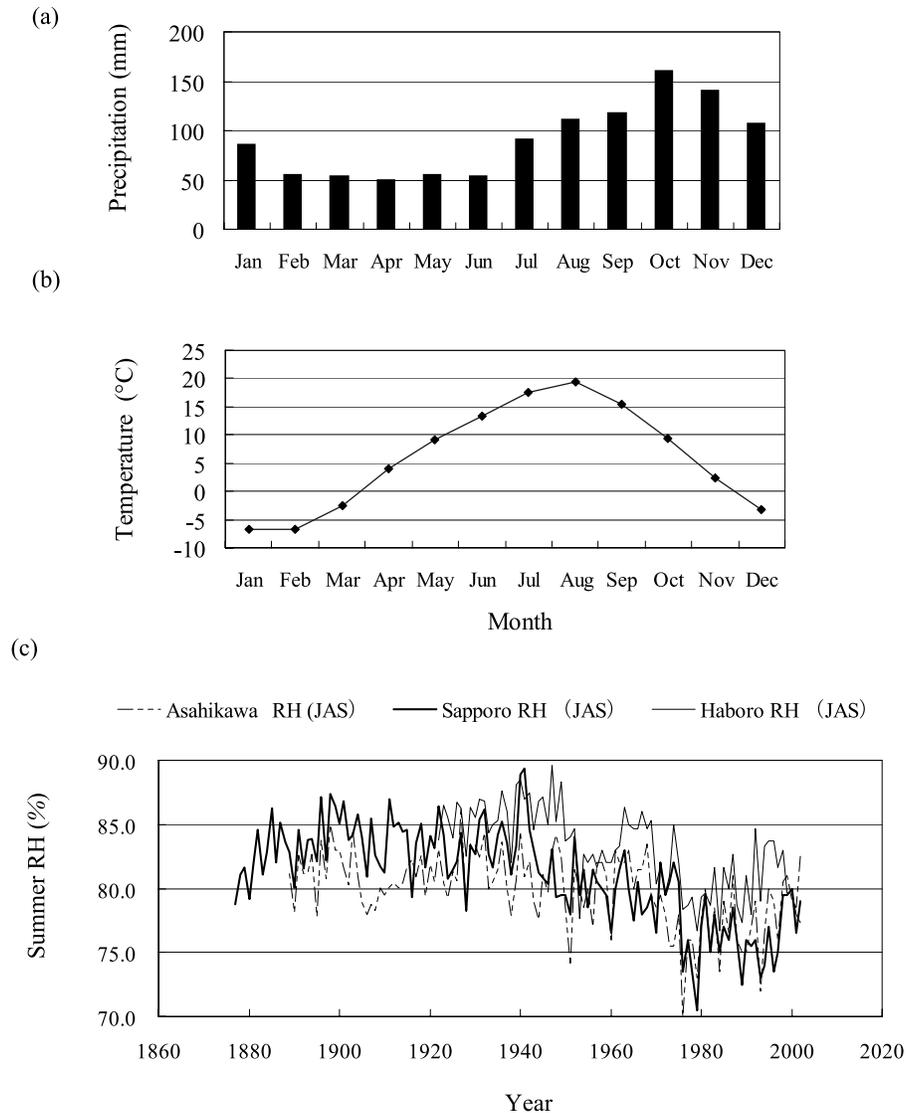


Figure 2. (a) The averages of monthly precipitation and (b) monthly temperatures from 1980 to 2002. (c) The interannual variation in the July–September RH at Sapporo (1877–2002), Asahikawa (1889–2002), and Haboro (1923–2002).

dendrochronological indices of the summer AO [*D'Arrigo et al.*, 2003] and annual PDO [*D'Arrigo et al.*, 2001; *Biondi et al.*, 2001; *MacDonald and Case*, 2005] in order to understand the relationship between the hydroclimate in northern Japan and the large-scale atmospheric circulation.

2. Materials and Methods

2.1. Meteorological Conditions of the Sample Site

[7] The sampling site (45°03'N, 142°06'E, 66 m asl) is located in a conifer-hardwood mixed forest in the Teshio Experimental Forest of Hokkaido University in northern Hokkaido (Figure 1). Hokkaido Island is the northernmost island in the Japanese archipelago and is surrounded by the Sea of Okhotsk, the Pacific Ocean, and the Sea of Japan (Figure 1). The precipitation over the Hokkaido Island mainly comprises snowfall during the winter monsoon period and rainfall from spring to fall associated with the Pacific polar front and cyclonic activity (extratropical cyclo-

nes and typhoon). In winter, the prevailing wind direction is from the northwest (i.e., from the Asian continent). The dry continental winds capture a large amount of moisture from the Sea of Japan by kinetic evaporation. Most of this moisture precipitates in the form of heavy snowfall on the northwestern side of the Japanese islands. On the other hand, in summer, the source of precipitation is the Pacific Ocean.

[8] Monthly climate data (precipitation and mean temperature) obtained at the nearest meteorological station (Toyotomi Meteorological Station (45°06.1'N, 141°46.7' E, 12 m asl)) around the sampling site averaged from 1980 to 2002 are shown in Figure 2. The maximum and minimum of monthly mean temperature are approximately 19.3 °C in August and -6.8 °C in February, respectively (Figure 2b). Monthly precipitation was maximum and minimum in October and April, respectively (Figure 2a). The annual precipitation amount averaged 1092 mm.

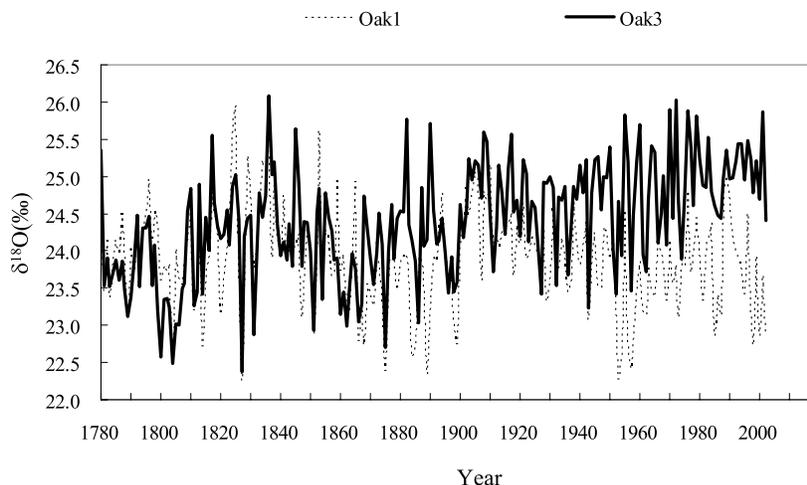


Figure 3. The $\delta^{18}\text{O}$ values of oak1 and oak3 during the entire period of 1776–2002. The correlation between the $\delta^{18}\text{O}$ values of oak1 and oak3 during the entire period 1776–2002 is significantly positive but weak ($r = 0.39$, $P < 0.0001$).

2.2. Sampling of Tree Rings

[9] In June 2003, several tree ring disks of oak were cut from the stumps remaining at the site of the “Carbon Cycle and Larch Growth experiment,” where all the trees in an area of 14 ha had been felled in the previous winter in order to investigate the effects of felling and planting on the carbon cycle in forest environments (refer to web site http://www.asiaflux.net/network/009TSE_1.html). These tree ring disks were stored and dried under room temperature.

[10] Two oak trees used in this isotopic investigation were growing at the site within a distance of approximately 100 m of each other. We reconstructed the summer RH at the sampling site using the oxygen isotope ratios of the two tree ring disks (oak1 and oak3) aged ~ 270 and ~ 230 years, respectively.

2.3. Isotope Analyses of the Tree Rings

[11] We analyzed oxygen isotopic ratios in the annual time resolution. A whole ring composed of earlywood and latewood was applied for isotopic analysis in this study although earlywood was usually removed in previous isotopic studies on oak trees to exclude the influence of carry-over material from the previous year [e.g., Robertson *et al.*, 2001; Raffalli-Delerce *et al.*, 2004; Danis *et al.*, 2006]. Hill *et al.* [1995] pointed out that when they measured isotopic ratios in earlywood and latewood, separately, the variation of carbon isotopic ratios in earlywood was similar to that of latewood in previous year. However, at the same time, the variation of oxygen isotopic ratios in earlywood was similar to that in latewood in current year. Our preliminary study has also suggested that the boundary between earlywood and latewood did not always mean the exact boundary of carry-over from the previous year. Therefore, we did not separate latewood from earlywood in cutting the annual ring.

[12] A block of each year ring was cut out by razor blade from 1773 to 2002 and 1776–2002 in oak 1 and oak 3, respectively. $20\mu\text{m}$ thick pieces of each ring were cut by a rotary microtome along edge-grained surface of the year ring block so that each piece represents the whole of the

year ring, and applied for the extraction of cellulose as described below. First, lipid was removed from each sliced ring by acetone and mixed solution of methanol and toluene. Then, lignin was decomposed by sodium chlorite solution including acetic acid. Finally, hemicellulose is removed by concentrated sodium hydroxide solution. The process of cellulose extraction was based on the revised method of Loader *et al.* [2003].

[13] The extracted α -cellulose from each sample was analyzed for the oxygen isotope ratio by using the continuous flow system of a pyrolysis-type elemental analyzer (ThermoQuest TCEA) and an isotope-ratio mass spectrometer (ThermoQuest Delta plus XL) [Sharp *et al.*, 2001]. In the pyrolysis furnace packed with glassy carbon, the oxygen in the cellulose is quickly converted to CO gas at 1375°C . Thus the isotopic ratios of CO gas can be analyzed for oxygen isotope ratios of the cellulose. The isotope ratio is expressed in the delta notation, i.e., $\delta^{18}\text{O} = ((^{18}\text{O}/^{16}\text{O})_{\text{sample}} / (^{18}\text{O}/^{16}\text{O})_{\text{standard}} - 1) \times 1000$ (‰), relative to the international standard (VSMOW) for oxygen isotopes. The standard deviation for the repeated analysis of a cellulose standard was within 0.3‰.

2.4. Meteorological Data

[14] We have investigated the correlations between the $\delta^{18}\text{O}$ in oak tree ring cellulose and climate parameters in Hokkaido Island in detail and found that the summer RH has the highest (negative) correlation coefficient with $\delta^{18}\text{O}$ of oak tree ring cellulose [Tsuji *et al.*, 2006]. The summer (July–August) RH data were averaged from among five meteorological observatories (Wakkanai ($45^\circ 24.9'\text{N}$, $141^\circ 40.7'\text{E}$, 3 m asl), Haboro ($44^\circ 21.7'\text{N}$, $141^\circ 42.0'\text{E}$, 8 m asl), Asahikawa ($43^\circ 45.4'\text{N}$, $142^\circ 22.3'\text{E}$, 120 m asl), Iwamizawa ($43^\circ 12.6'\text{N}$, $141^\circ 47.1'\text{E}$, 42 m asl), and Sapporo ($43^\circ 03.5'\text{N}$, $141^\circ 19.7'\text{E}$, 17 m asl)) in the same climatic division as the sampling site during the 1961–2002 period [Tsuji *et al.*, 2006]. Therefore, the oxygen isotope ratios of the oak tree ring cellulose in this forest can become a proxy of summer RH. However, the period of 42 years (1961–2002) is too short to perform the calibration and verification

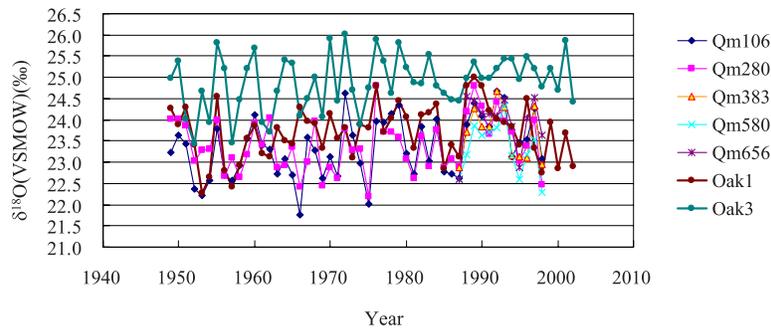


Figure 4. The $\delta^{18}\text{O}$ values of oak1 and oak3 during 1953–2002 together with those of Qm106 and 280 during 1949–1998 and Qm383, 580, and 656 during 1987–1998 in Uryu site [Nakatsuka *et al.*, 2004].

between the oxygen isotope ratios of the tree ring cellulose and the observed summer RH data in order to reconstruct the past summer RH. In general, observed climate data for more than 100 years is ideally required for the calibration and verification. The RH data for more than 100 years are available at only two sites, Sapporo (1877–2002) and Asahikawa (1889–2002), among the five meteorological observatories. Both the summer RH data sets indicate decreasing trends (Figure 2c). If the decreasing trend of the summer RH is caused by local aridification due to the progress of urbanization at Asahikawa and Sapporo in the twentieth century, the historical changes in the summer RH at the sampling site in the natural forest could be different from those in Sapporo and Asahikawa. However, this decreasing trend is also seen in Haboro (1923–2002) (Figure 2c), which has the third longest span of meteorological data among the five observatories and is located on the coast of the Sea of Japan where it is less affected by local urbanization than Sapporo and Asahikawa. This shows that the decreasing trend of the summer RH in Asahikawa and Sapporo may be mainly due to the change in climate across the whole of northwestern Hokkaido. We used the averaged RH at Sapporo and Asahikawa because by averaging the time series of two or more meteorological station records, many problems associated with the heterogeneities of records due to the differences in the station microclimates may be avoided, thereby increasing the reliability of the data in order to calibrate the tree ring chronologies [Blasing *et al.*, 1981; Jacoby *et al.*, 2000; Singh and Yadav, 2005].

3. Results and Discussion

3.1. Interindividual Correlations

[15] Historical fluctuations of the $\delta^{18}\text{O}$ values of tree ring cellulose of oak1 and oak3 during the entire period of 1776–2002 are shown in Figure 3. Although the high-frequency fluctuations in the $\delta^{18}\text{O}$ values of oak1 and oak3 correspond to each other, the long-term trends of the $\delta^{18}\text{O}$ values are different during the entire period. The variations in the $\delta^{18}\text{O}$ values of oak3 exhibit a long-term increasing trend in opposition to the variations in oak1, which does not exhibit such a trend. These inconsistent trends cause a significantly positive but weak correlation between the $\delta^{18}\text{O}$ values of oak1 and oak3 during the entire period of 1776–2002 ($r = 0.39$, $P < 0.0001$) (Figure 3). Because the $\delta^{18}\text{O}$ values of oak1 and oak3 are negatively correlated with the summer RH [Tsuji *et al.*, 2006], the decreasing trend in

the summer RH over the whole of northwestern Hokkaido should be reflected by the increasing trend in the $\delta^{18}\text{O}$ values of the tree ring cellulose. This is true for the case of oak3 alone. The long-term trend of oak1 may be countered by some other effects. For example, this may be due to the vertical distribution of the soil water taken up by the roots of oak1. With an increase in the tree age, oak1 may preferentially absorb deeper soil water originating from melted snow with low $\delta^{18}\text{O}$ values rather than shallow soil water from summer precipitation with high $\delta^{18}\text{O}$ values. Therefore, the long-term summer aridification might be then canceled because of the long-term change in the soil-water $\delta^{18}\text{O}$ in the case of oak 1. These differences in responses between the $\delta^{18}\text{O}$ values of oak1 and oak3 are possibly due to the heterogeneous environments of the soil water. Because the inconsistent trends of the $\delta^{18}\text{O}$ values of oak1 and oak3 with respect to the soil water provide a weak individual correlation, we cannot use the low-frequency $\delta^{18}\text{O}$ variations of the oak including the local hydrological effect for a climatological discussion.

[16] On the other hand, the high-frequency $\delta^{18}\text{O}$ variations in oak 1 and 3 may hold the signal of regional climate because they are similar to each other. In order to certify that the high-frequency $\delta^{18}\text{O}$ variations represent some regional climatic signals, we compared the $\delta^{18}\text{O}$ values of oak1 and oak3 to those of the same species in the Uryu Experimental Forest of Hokkaido University in northern Hokkaido during 1949–1998 (Figure 4), which were previously reported by Nakatsuka *et al.* [2004]. Uryu site is at about 70 km distance southward from the sampling site (Figure 1). The climate of Uryu site is similar to that of the sampling site. Table 1 indicates that the variations in the $\delta^{18}\text{O}$ values of oak1 and oak3 are correlated well with those of Qm106, Qm280, Qm383, Qm580, and Qm656 in Uryu site during 1949–1998 and 1987–1998 [Nakatsuka *et al.*, 2004]. The variation in the average $\delta^{18}\text{O}$ values of oak1 and oak3 is highly correlated with that of Qm106 and Qm280 during 1949–1998 (correlation coefficient = 0.62, $P = 2.0 \times 10^{-6}$, Table 1). These high correlations for the 12 and 50 years suggest that in the case of this short period, the variation in $\delta^{18}\text{O}$ values of oak1 and oak3 is controlled by regional climate without effect of long-term change due to the local effect such as hydrological setting.

[17] Hereafter, in order to focus on the high-frequency $\delta^{18}\text{O}$ variation, we eliminated the low-frequency components of more than 250 years from both the $\delta^{18}\text{O}$ time series of oak 1 and 3 by using a high-pass filter. Then, the high-pass

Table 1. Correlation Coefficients Between Variations in the $\delta^{18}\text{O}$ of Oak1 and Oak3 in This Study and Qm106, Qm280, Qm383, Qm580, and Qm656 in Uryu Site Together With Those Between the Averaged $\delta^{18}\text{O}$ Values of Oak1 and Oak3 in This Study and Qm106 and Qm280 in Uryu Site^a

	Qm280	Qm106	Qm383	Qm580	Qm656	Average (Qm106 and Qm280)
Oak1	0.54***	0.53***	0.41	0.60**	0.60**	
Oak3	0.36*	0.55***	0.45	0.62**	0.65**	
Average (oak1 and oak3)						0.62***

^aUryu site is from Nakatsuka *et al.* [2004]. The analyzed periods of Qm383, Qm580, and Qm656 are from 1987 to 1998 and those of oak1 and oak3 are from 1949 to 1998. Single, double, and triple asterisks show the probabilities of $P < 0.1$, $P < 0.05$, and $P < 0.001$, respectively. The bold and italic numbers mean periods of comparison (50 and 12 years), respectively.

filtered $\delta^{18}\text{O}$ values were normalized during 1776–2002 in order to discuss the variation rather than the absolute value. “Normalize” means that deviation of $\delta^{18}\text{O}$ in each year from the entire mean value during 1776–2002 was divided by the entire standard deviation. The high-pass filtered and normalized $\delta^{18}\text{O}$ values of oak1 and oak3 are shown in Figure 5. The annual-to-several-years scale and decadal-to-centennial scale fluctuations in the $\delta^{18}\text{O}$ values of oak1 and oak3 now correspond better to each other. Indeed, the interindividual correlation between the $\delta^{18}\text{O}$ values of oak1 and oak3 becomes stronger ($r = 0.56$, $P < 0.0001$) during 1776–2002 (Figure 5). This high correlation confirms that high-pass filter removes effect of local hydroenvironment and the high-pass filtered $\delta^{18}\text{O}$ values of oak1 and oak3 becomes a proxy of regional climate.

3.2. Relationship Between $\delta^{18}\text{O}$ of Tree Ring Cellulose of the Oak and the Summer RH

[18] We have already found that the interannual variations in the $\delta^{18}\text{O}$ values in the oak at the sampling site are sensitive to the July–August RH, which is averaged among the data from five observatories (Sapporo, Asahikawa, Haboro, Iwamizawa, and Wakkanai, Figure 1) during

1961–2002 [Tsuji *et al.*, 2006]. However, since we averaged the climatic parameters at two observatories (Sapporo and Asahikawa) here for calibrating the $\delta^{18}\text{O}$ values in the oak during 1889–2002, we reexamined the correlation between the variations in the $\delta^{18}\text{O}$ values of the oak and the averaged climatic parameters (temperature, precipitation, and RH) which have been instrumentally observed in Sapporo and Asahikawa. In order to focus on the short-term variations, the climatic parameters were also high-pass filtered against frequencies greater than 250 years and normalized as well as that in $\delta^{18}\text{O}$ value. The best correlation was found with the averaged RH in July–September (summer) during 1889–2002. The summer RH and the $\delta^{18}\text{O}$ of oak are shown in Figure 5 (the axis of summer RH is reversed). The variations in the summer RH and the $\delta^{18}\text{O}$ of oak appear to correspond throughout the examined period (Figure 6). We compared the regression expressions and the correlation coefficients among 1889–2002, 1889–1941, and 1941–2002 (Table 2a). The regression expressions during 1889–2002, 1889–1941, and 1941–2002 are almost similar (Table 2a and Figure 6). The correlation coefficients are -0.515 ($P < 0.0001$) during 1889–2002, -0.521 ($P = 0.0007$) during 1889–1940, and -0.520 ($P = 0.0001$) during 1941–2002. When the periods of calibration and verification are 1889–1940 and 1941–2002, respectively, the coefficient of efficiency (CE) is 0.25 (Table 2b). Further, when the periods of calibration and verification are 1941–2002 and 1889–1940, respectively, CE is 0.21 (Table 2b). Because both the CEs are above 0 [Cook *et al.*, 1999] and both the regression expressions are almost the same during 1889–1941 and 1941–2002, we can conclude that the negative correlation between the interannual variation in the $\delta^{18}\text{O}$ and the summer RH is applicable to the period before which the instrument data were obtained. The relationship between the interannual variation in the $\delta^{18}\text{O}$ and the summer RH is almost constant from 1889 to 2002. Therefore, we used the regression expression during 1889–2002 to reconstruct the summer RH extending back into the past up to 1776. However, because we use the detrended data, our discussion is limited to variations with frequencies

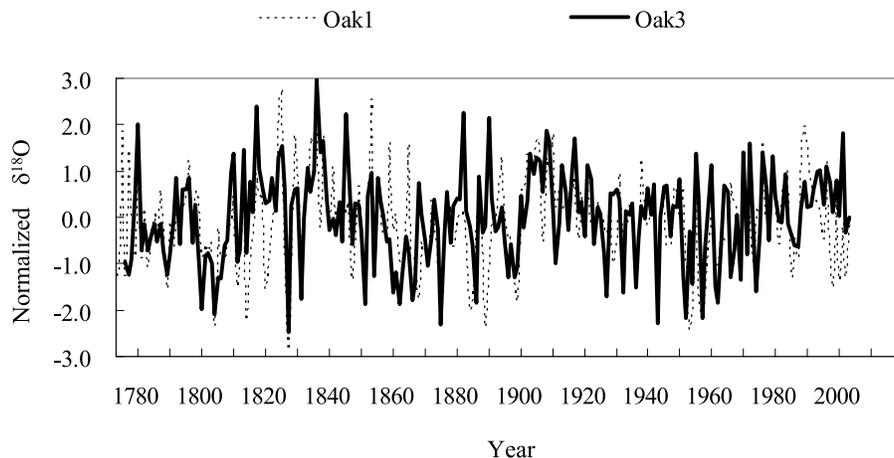


Figure 5. The interannual changes in the detrended $\delta^{18}\text{O}$ values of oak1 and oak3. Variations with periods greater than 250 years are eliminated with high-pass filters (detrend).

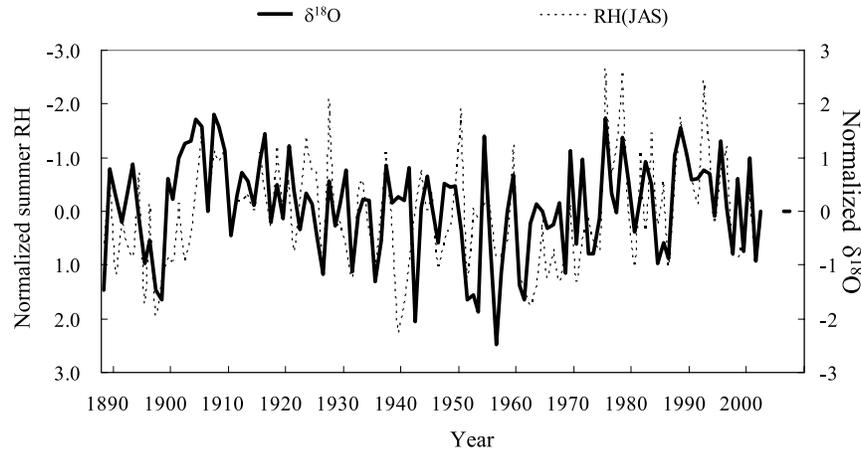


Figure 6. Detrended July–September RH and the averaged $\delta^{18}\text{O}$ anomalies of oak1 and oak3 during 1776–2002. Variations in periods greater than 250 years are eliminated with high-pass filters (detrend). Detrended data are normalized. The y axis of the summer RH anomaly is reversed.

less than the 250-year frequency of the reconstructed summer RH.

3.3. Reconstructed Summer RH and Its Spectral Analysis

[19] On the basis of the regression expression between the interannual changes in the normalized $\delta^{18}\text{O}$ of oak and the normalized RH in July–September during 1889–2002, we reconstructed the normalized summer RH from 1776 to 2002, in which the trends longer than the 250-year frequency were removed (Figure 7). In the first half of the nineteenth century (standard deviation = 0.62), the variability of the summer RH is greater than that in the second half of nineteenth century and in the twentieth century (Standard deviation = 0.53). The periods of high RH are 1797–1812, 1857–1894 and 1950s. The periods of low RH during 1831–1836 (Tempo famine in Japan) and in the 1900s correspond to the historical evidences of the famines in northeastern Japan.

[20] We analyzed the time series of the reconstructed summer RH using the multitaper spectral method (MTM) in order to extract the prominent cycle [Mann and Lees, 1996]. The power spectrum is shown in Figure 7. The broken lines indicate 90, 95, and 99% levels of significance. The significant peaks are found in the ranges of 50–100+ years, 12.9 years, 4–5 years, and 2–3 years. The peaks at 2–3 years and 4–5 years are identified as the monsoon cycle [Li and Zhang, 2002] and ENSO (El Niño–Southern

Oscillation) cycle, respectively. The ~ 12.9 -year peak may be related to solar effects, which can influence the climate through the AO forcing [Shindell et al., 2001]. The broad 50–100+ year peak that includes a ~ 70 -year oscillation was observed in the global instrumental record [Mann and Park, 1994; Schlesinger and Ramankutty, 1994].

[21] As a result of spectral analysis, the variation of the reconstructed RH was found to be influenced by the various teleconnection (ENSO, PDO and AO etc.) There were several past studies on the teleconnection between dendrochronologically reconstructed climate variations at remote areas. For example, Norman and Taylor [2003] shows teleconnection between the reconstructed forest fire in California, North America and the reconstructed PDO and/or ENSO. Schongart et al. [2004] illustrated teleconnection between the reconstructed flood in Amazon and the reconstructed El Niño events. D’Arrigo et al. [2005] indicated teleconnection between the reconstructed Siberian high index and the reconstructed ENSO. Such previous teleconnection studies have mainly investigated on interannual scale. In this study, the time series of reconstructed summer RH in northern Hokkaido shows its most prominent power in the range of 50–100+ years (Figure 8). Therefore, we focus on the teleconnection on decadal-centennial scales. 50–100+ year cycle is found in the summer AO [D’Arrigo et al., 2003] and annual PDO indices [D’Arrigo et al., 2001] reconstructed from tree ring studies in the high- and middle-latitude regions in the

Table 2a. Single Regressions Between the Detrended $\delta^{18}\text{O}$ and the July–September RH During 1889–2002, 1889–1941, and 1941–2002^a

Periods	Multiple Regression	Correlation Coefficient	P
1889–2002	$\text{RH} = (-0.56 \pm 0.09) \times \delta^{18}\text{O} - 0.046 \pm 0.080$	-0.515	<0.0001
1941–2002	$\text{RH} = (-0.57 \pm 0.12) \times \delta^{18}\text{O} - 0.021 \pm 0.012$	-0.520	0.0001
1889–1940	$\text{RH} = (-0.56 \pm 0.13) \times \delta^{18}\text{O} + 0.106 \pm 0.060$	-0.521	0.0007

^aCorrelation coefficients and P values of each equation are shown. $\delta^{18}\text{O}$ is the averaged oxygen isotope ratios in the tree ring cellulose of oak1 and oak3 during 1889–2002. The July–September RH is the averaged RH at Sapporo and Asahikawa during 1889–2002.

Table 2b. Coefficient of Efficiency (CE) and Correlation Coefficient Considered for the Calibration and Verification Between the Detrended $\delta^{18}\text{O}$ Anomaly and July–September RH During 1889–1940 and 1941–2002 and 1941–2002 and 1889–1940^a

Calibration Periods (Year)	Verification Periods (Year)	CE
1889–1940	1941–2002	0.25
1941–2002	1889–1940	0.21

^a $\delta^{18}\text{O}$ is the averaged oxygen isotope ratios in the tree ring cellulose of oak1 and oak3 during 1889–2002. The July–September RH is the averaged RH at Sapporo and Asahikawa during 1889–2002.

Northern Hemisphere. Thus the variations on decadal-centennial scale in the summer hydroclimate in Hokkaido, northern Japan, might relate to the PDO and AO. Next, we discuss the relationships between the variation on decadal-centennial scale in the summer RH and the atmospheric indices (PDO and AO).

3.4. Relationship Between Summer RH and Annual PDO Index

[22] The PDO is a recurring pattern of ocean-atmosphere variability in which the central gyre and western area are cool at the same time the eastern margin becomes warm or vice versa. The alternating phases of the PDO can last for two to three decades with reversals being noted in 1924/1925, 1946/1947, and 1976/1977 [Mantua *et al.*, 1997]. The PDO forcing strongly influences the midlatitude zone of the coastal area of the North Pacific Ocean. The instrumentally observed PDO index is the leading empirical orthogonal function (EOF) of the SST anomaly in the North Pacific Ocean, poleward of 20°N [Mantua *et al.*, 1997] and the record was obtained from a web site (ftp://ftp.atmos.washington.edu/mantua/pnw_impacts/INDICES/PDO.latest by N. Mantua).

[23] We investigated the relationship between the decadal-centennial scale variations in the summer RH in Hokkaido, northern Japan, and the PDO indices. In order to demonstrate the decadal-centennial scale variations, both of the RH and PDO indices (reconstructed and instrumentally observed) are high-pass filtered against longer than 250 years periodicity, then normalized and averaged at 11-year running intervals. We calculated the correlation coefficients between the variations with periodicities ranging from 11 to 250 years in the reconstructed summer RH in this study and the annual PDO indices previously reconstructed from tree ring widths in North America during 1781–1974 [D'Arrigo *et al.*, 2001], 1781–1986 [Biondi *et al.*, 2001] and 1781–1991 [MacDonald and Case, 2005]. We also examined the correlation between the instrumentally observed summer RH in northwest Hokkaido, Japan, and the annual PDO index during 1905–1997 by Mantua *et al.* [1997]. The variations in the 11- to 250-year periodicities in the instrumentally observed summer RH are negatively correlated with those of the observed annual PDO index during 1905–1997 ($r = -0.34$, $P = 0.0085$) (Figure 9b). The variations in the 11–250 year periodicities in the reconstructed summer RH in this study are negatively correlated with the previously reconstructed annual PDO indices ($r = -0.43$, $P < 0.00001$ [D'Arrigo *et al.*, 2001], $r = -0.34$, $P < 0.00001$ [Biondi *et al.*, 2001] and $r = -0.34$, $P < 0.00001$ [MacDonald and Case, 2005], Figure 9a). On Hokkaido Island, northern Japan, the climate conditions become wet when the PDO indices are low and vice versa.

[24] In order to clarify the relationship between the hydroclimate in Hokkaido and the PDO, we examined the spatial distribution of the correlation coefficients between the annual changes. The July–September sea level pressure field and the instrumentally observed annual PDO index during 1948–2002 were obtained from the reanalysis data from the National Center for Environment Predictions/the National

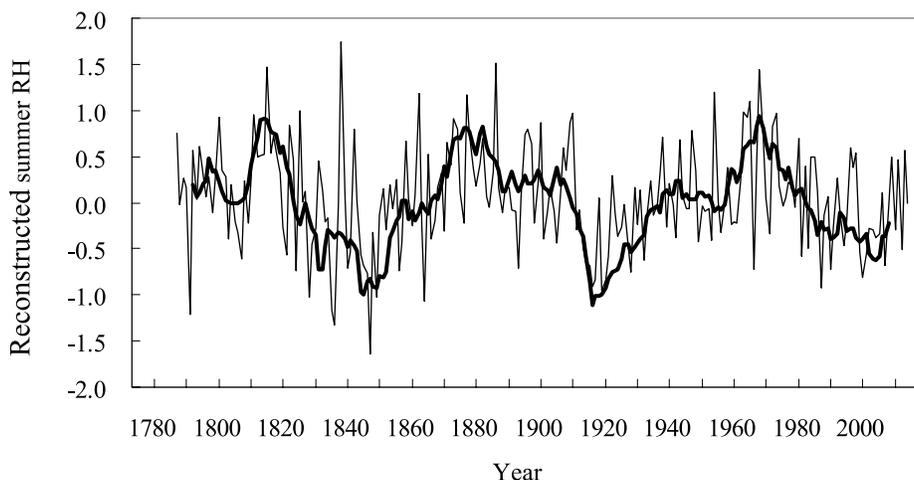


Figure 7. Interannual changes and the 11-year running average of the reconstructed summer RH during 1776–2002. On the basis of the regression expressions between the interannual changes in the $\delta^{18}\text{O}$ of oak and the RH in July–September during 1889–2002, the summer RH anomaly during 1776–2002 is reconstructed and shown (bold line, interannual; fine line, 11-year running average). These data are detrended and normalized.

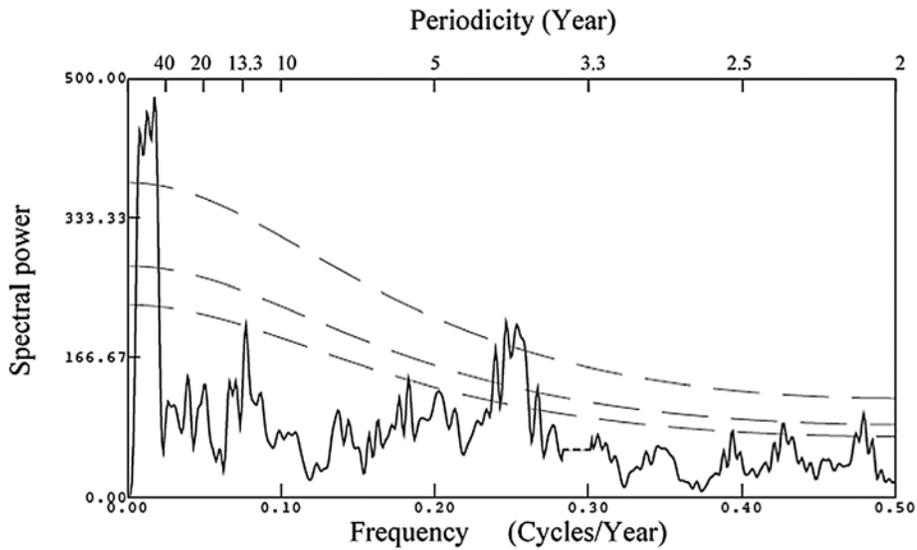


Figure 8. The power spectrum analyzed by the Multitaper method (MTM) of spectral analysis for the reconstructed RH. Dashed lines indicate 90, 95, and 99% levels of significance.

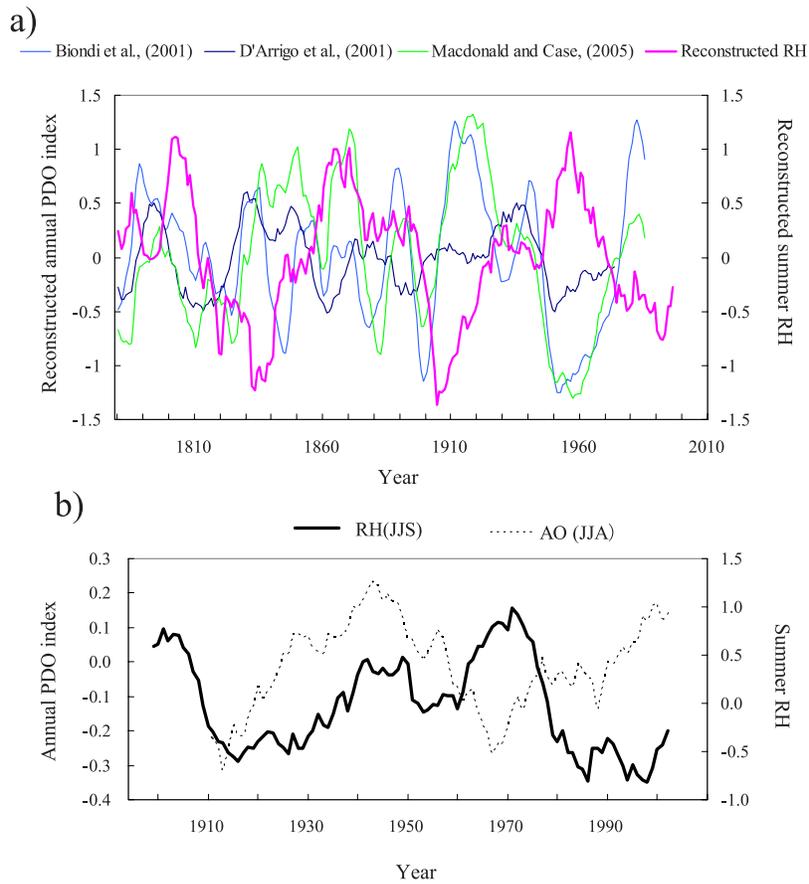


Figure 9. (a) The variations in 11- to 250-year periodicities of the reconstructed summer RH and annual PDO indices [Biondi et al., 2001; D'Arrigo et al., 2001; MacDonald and Case, 2005] during 1781–1974. These data are detrended. (b) The variations in 11- to 250-year periodicities in the observed summer RH and observed annual PDO index during 1905–1997.

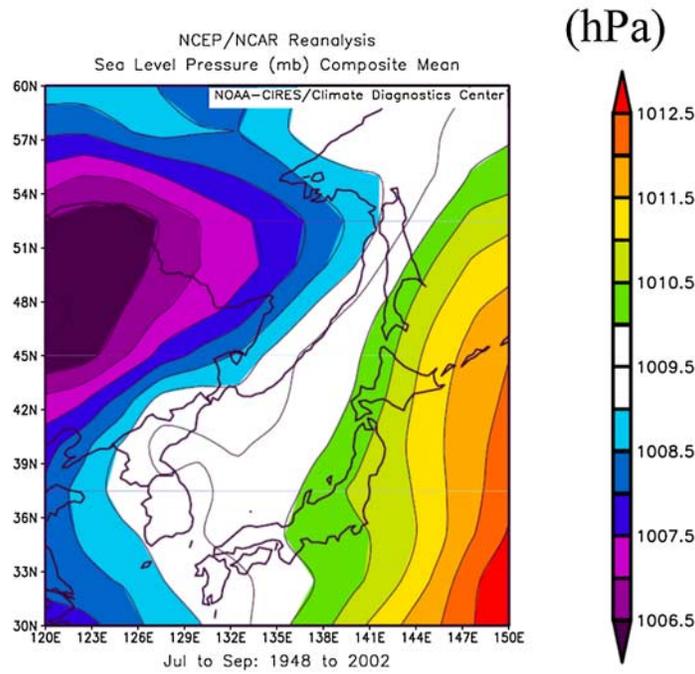


Figure 10. The average July–September sea level pressure field around Japan during 1948–2002. The sea level pressure data of the reanalysis data from NCEP/NCAR were used.

Centers for Atmospheric Research (NCEP/NCAR), and the 11- to 250-year periodicities in these data were not removed.

[25] First, the average July–September sea level pressure field around Japan during 1948–2002 shows that the hydrological climate of Japan in summer is usually influenced by the southerly humid wind, namely, the summer monsoon, due to the interaction between the Pacific high and the Asian continental low (Figure 10). Next, the spatial distribution of the correlation coefficients around Japan during 1948–2002 (Figure 11) shows that when the PDO

index is lower, the Pacific high and the Asian continental low in summer intensify and the southerly humid wind blows strongly and the summer RH increases in northern Japan and vice versa. Indeed, this relationship corresponds to the negative correlation between the interannual changes of July–September RH in northwestern Hokkaido and the annual PDO during 1948–2002 ($r = -0.41, P = 0.002$). The negative relationship between the variations in the 11- to 250-year periodicities of the summer RH and the annual PDO index is also consistent with the result of the spatial

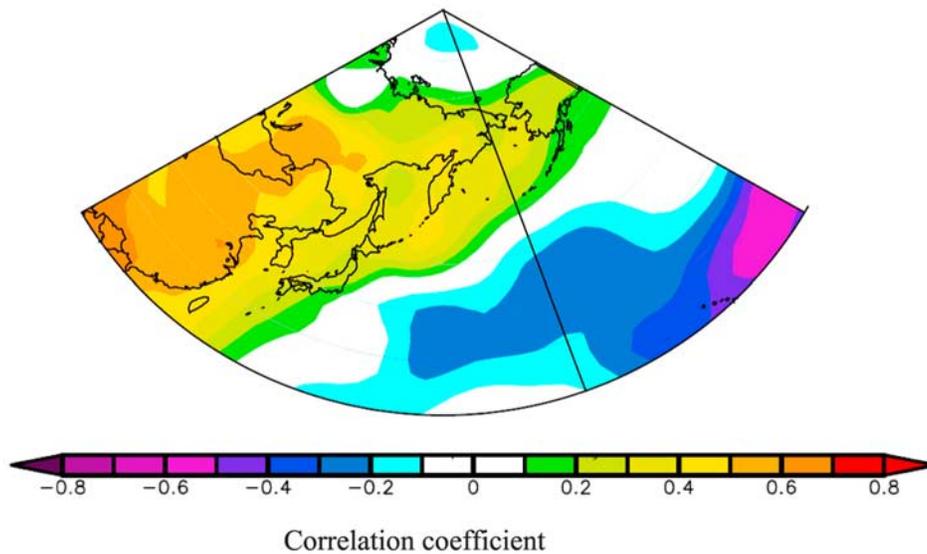


Figure 11. The horizontal distribution of the correlation coefficients between the interannual variations of July–September sea level pressure and the July–September PDO index (1948–2002) around Japan. The sea level pressure data of the reanalysis data from NCEP/NCAR were used.

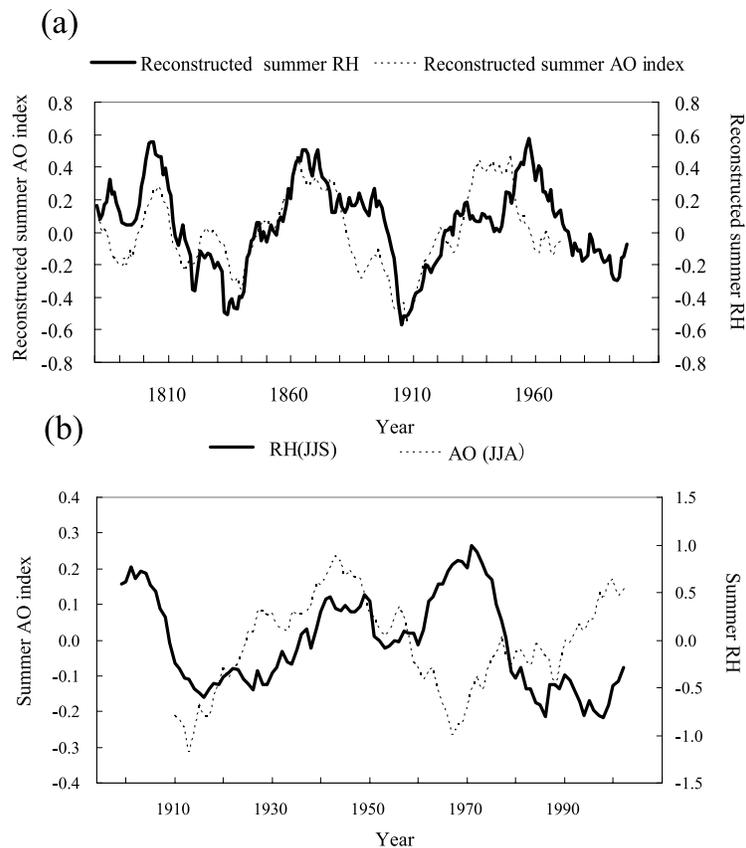


Figure 12. (a) The variations in 11- to 250-year periodicities in the reconstructed summer RH and reconstructed summer AO index [D'Arrigo *et al.*, 2003] during 1780–1930. (b) The variations in 11- to 250-year periodicities in the observed summer RH and summer AO index during the period 1940–1997.

distribution of the correlation coefficients around Japan. We can conclude that the variations in the 11- to 250-year periodicities of the hydroclimate in Hokkaido, northern Japan, is consistently influenced by the PDO through the intensities of the Pacific high and the Asian continental low during 1781–2002.

[26] Shen *et al.* [2006] showed that during the period 1951–2000, the positive phases of the annual PDO coincided with anomalous dry periods in northern and southern China and wet periods in middle China in the Yangtze River Valley owing to a weakened Pacific high; this shows the dipole pattern. This negative correlation in north China (north of 40°N), whose latitude is similar to that of Hokkaido ($45^{\circ}03'\text{N}$, the sampling site), corresponds to that between the summer RH in Hokkaido and the annual PDO index. The relationship between the hydrological climate and the PDO may range over east Asia on a synoptic scale. On the basis of the relationship between the instrumentally observed PDO index and a data set of the drought/flood indices (D/F) in east China during 1925–1998, Shen *et al.* [2006] reconstructed the PDO index from a proxy of the D/F since 1470. This reconstruction presumes that the water cycle in east Asia is derived by the PDO alone. However, Gong and Ho [2003] indicated that the water cycle in east Asia is also related to the AO index. Hence, we can expect that the water cycle in east Asia is derived not only by the PDO but also by the AO. In order to verify the relationship

between the water cycle and the AO, we investigated the correlation between the summer RH and the AO index.

3.5. Relationship Between Summer RH and Summer AO Index

[27] The arctic Oscillation (AO) was introduced as a mode of atmospheric circulation by Thompson and Wallace [1998]. This primary mode of dynamics in the atmosphere predominates the extratropical Northern Hemisphere circulation from the surface to the lower stratosphere during cold seasons (November–April) [Thompson and Wallace, 2000]. Although many researchers have discussed the strong influences of the winter AO on the large-scale atmospheric circulation, the AO also explains a major portion of the total variance in the atmospheric circulation during warm seasons (May–October) [Thompson and Wallace, 2000]. The variations in the AO are characterized by a seesaw pattern in which the atmospheric pressure and mass in the northern polar and midlatitudes alternate between positive and negative phases [Wallace, 2000]. The observed AO indices are determined by the time series of the leading principal component of the monthly mean sea level pressure in the Northern Hemisphere (poleward of 20°N) [Thompson and Wallace, 1998]. These indices have an instrumental time series for 1899–2001 [Thompson and Wallace, 2000]. The data are prepared and provided on the Web site of David W. J. Thompson (http://www.atmos.colostate.edu/ao/Data/ao_index.html).

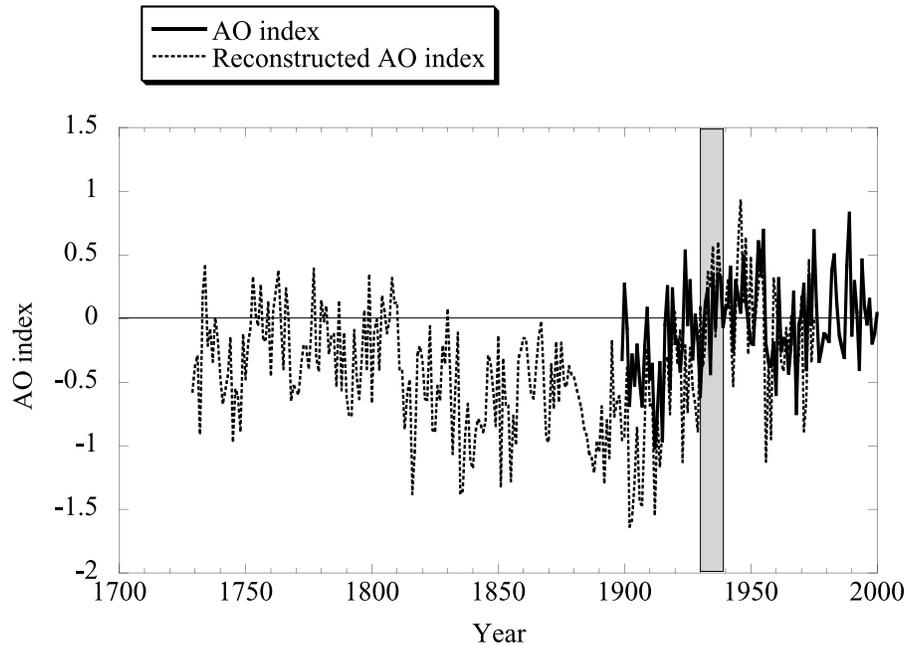


Figure 13. Observed summer AO index (1899–2002) and reconstructed summer AO index [D’Arrigo *et al.*, 2003] (1776–2002). Shaded area represents the 1930s. These are original data.

[28] We examined the relationship between the decadal-centennial-scale variations in the summer RH in northern Japan and the AO. In order to demonstrate the decadal-centennial-scale variations, the AO indices (reconstructed and instrumentally observed) are high-pass filtered against longer than 250 years periodicity, then normalized and averaged at 11-year running intervals as well as those of PDO indices. We compared the variations in the 11–250 years periodicities in the reconstructed summer (July–September) RH with that of the dendrochronologically reconstructed summer (June–August) AO index [D’Arrigo *et al.*, 2003] (AO SLP index). We also compared the variations in the 11–250 years periodicities in the instrumentally observed summer (July–September) RH with those of the instrumentally observed summer AO index (Figure 12b). The relationships between the summer RH and AO index probably change during the 1930s (Figures 12a and 12b). In order to investigate this change in relationship in detail, we calculated the correlation coefficients between the summer RH and summer AO index before and after the 1930s. The fluctuation in the 11- to 250-year periodicities in the reconstructed summer RH is positively correlated with that of the reconstructed summer AO index during 1781–1930 ($r = 0.74$, $P < 0.00001$). Because the period of the reconstructed summer AO after the 1930s was short (1940–1970), the correlation coefficient between the longer observed AO index and the summer RH during 1940–1997 was calculated. In contrast to the period of 1781–1930, the variations in the 11- to 250-year periodicities in summer RH are negatively correlated with those of the observed summer RH during 1940–1997 ($r = -0.58$, $P < 0.00001$) (Figure 12b). This result indicates that the relationship between the summer RH and the summer AO has clearly changes during the 1930s.

[29] We observed the “raw” AO index variations from which the low-frequency cycles of more than 250 years and

high-frequency cycles of less than 11 years are not removed. The fluctuations in the instrumentally “raw” observed summer AO index (1899–2002) and the “raw” reconstructed summer AO index [D’Arrigo *et al.*, 2003] (1776–2002) are shown in Figure 13. The 1930s period is when the summer AO index changes from negative to positive values (Figure 13). This corresponds to the period when the shift in relationship occurred from the decadal to centennial variations in the summer RH and summer AO. The shift from the negative to the positive AO index and/or global warming might change the relationship between the hydroclimate in northern Japan and the hemispheric atmosphere circulation.

[30] For the purpose of exploring whether the reversed relationship between the summer hydroclimate in northern Japan and the summer AO before and after the 1930s are caused by the shift in the summer AO index itself from negative to positive, we investigated the differences in atmospheric circulation patterns between the periods of the negative AO index (1899–1930) and the positive AO index (1970–1997). We examined the spatial distribution of the correlation coefficients between the interannual variations in the sea level pressure in July–September [Trenberth and Hurrell, 1994] and the July–September (summer) AO indices during 1970–1997 and 1899–1930 (Figures 14a and 14b). Figure 14a is similar to the recent summer AO pattern shown by Ogi *et al.* [2004]. However, Figure 14b is not similar to that AO pattern. During 1970–1997, the intensified low around northeast China clearly exists during the negative AO phase (Figure 14a). Because the summer (July–September) season in Hokkaido is basically dominated by a southerly humid wind (summer monsoon) due to the dominant Pacific high and the Asian continental low (Figure 10), the humid wind from the south weakens because of the weak low around northeast China with a positive AO phase during 1970–1997 (Figures 14a and 14b).

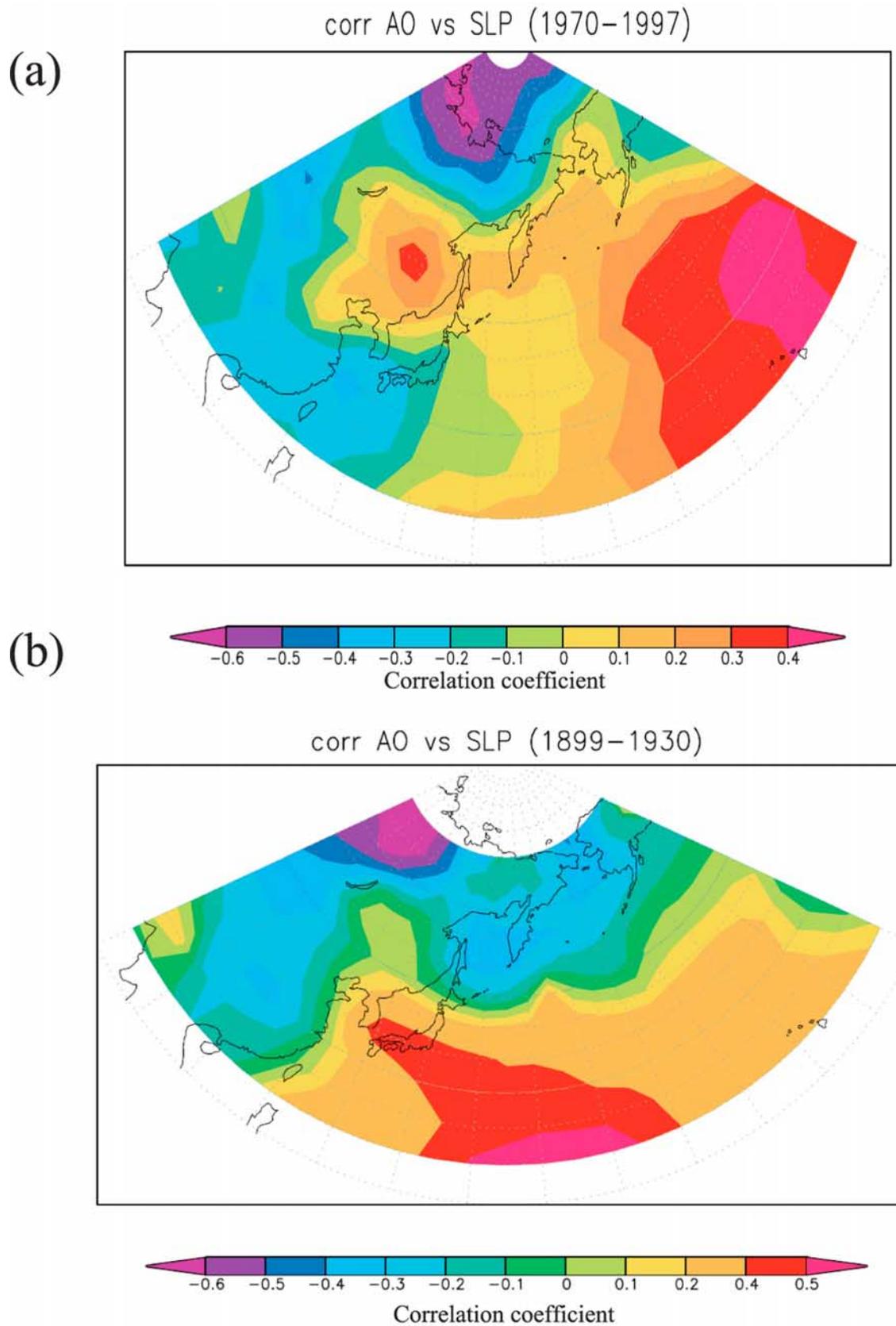


Figure 14. The horizontal distribution of the correlation coefficient between the interannual variations in the July–September sea level pressure [Trenberth and Hurrell, 1994] and July–September AO index during the periods (a) 1970–1997 and (b) 1899–1930.

Table 2c. Multiple and Single Regressions Between the Decadal-Centennial Changes in the Reconstructed (Observed) Summer RH and Indices of the Reconstructed Summer AO Index and/or Annual PDO During 1781–1930 (1940–1996)^a

Periods	Multiple and Single Regressions	R ²	P	N
1940–1996	RH = $(-0.41 \pm 0.06) \times \text{PDO} + (-0.42 \pm 0.15) \times \text{AO} - 0.05 \pm 0.06$	0.46	5.5×10^{-8}	57
1940–1996	RH = $(-0.53 \pm 0.09) \times \text{PDO} - 0.03 \pm 0.06$	0.38	3.4×10^{-7}	57
1940–1996	RH = $(-0.70 \pm 0.16) \times \text{AO} - 0.09 \pm 0.06$	0.27	3.9×10^{-5}	57
1781–1930	RH = $(-0.30 \pm 0.06) \times \text{PDO} + (0.96 \pm 0.07) \times \text{AO} - 0.04 \pm 0.03$	0.63	2.1×10^{-32}	150
1781–1930	RH = $(-0.47 \pm 0.08) \times \text{PDO} - 0.02 \pm 0.04$	0.17	1.1×10^{-7}	150
1781–1930	RH = $(1.04 \pm 0.08) \times \text{AO} + 0.04 \pm 0.03$	0.56	1.9×10^{-8}	150

^aThe R² and P values in each equation and the number of samples are shown.

Then, the summer RH drops with a positive AO phase. During 1899–1930, with a positive AO phase, the Pacific high simply becomes stronger and then a strong southerly wet wind (summer monsoon) should then blow toward Hokkaido. Moreover, the summer RH increases with the positive AO phase.

[31] The correlation coefficients between the summer AO and the sea level pressure are totally significantly low compared to the case of annual PDO (Figure 14a). This is because the active season for the AO is not summer but winter. Therefore, the annual variation in the AO in summer may not strongly affect the hydroclimate in northern Japan. However, the relationships between the summer RH and summer AO-related sea level pressure during 1899–1930 and 1970–1997 are consistent with the results of the reversed relationship between the decadal and above scale variations in the summer RH and AO index in summer before and after the 1930s. Thus, the sea level pressure related to the summer AO may control the hydroclimate in summer in northern Japan, and the transformed distributions of the sea level pressure due to the shift in the 1930s should provide a reversed relationship between the decadal-centennial-scale changes in the hydroclimate in northern Japan and the summer AO. Since the AO is related to the Northern Hemispheric temperature [Thompson and Wallace, 1998; D'Arrigo *et al.*, 2003], the increase in temperature since the beginning of twentieth century might provide a reversed relationship between more decadal changes in the hydroclimate in northern Japan and the summer AO.

3.6. Relationships Between Summer RH and Indices of Summer AO and Annual PDO

[32] The reconstructed summer RH in northwestern Hokkaido, Japan, is highly correlated with both the PDO and AO on the decadal-centennial scale. However, PDO and AO may be dependent on each other. Therefore, for example, the correlations between the summer RH and AO may be indirect and the only PDO is directly related to the summer RH. We conducted a multiple regression analysis of the high-pass filtered, normalized and 11-year running averaged time series of AO, PDO and RH in order to assess whether both of AO and PDO independently govern the summer RH in northern Hokkaido, Japan. If either AO or PDO does not control the RH directly, the partial regression coefficient of AO or PDO should show a high standard error. Here, we used the reconstructed PDO index of D'Arrigo *et al.* [2001], which shows the highest correlation with the reconstructed RH. The relationships between the variations in the 11- to 250-year periodicities in the summer RH and the indices of the summer AO and annual PDO are shown by multiple regression expressions in Table 2c. The

reconstructed and instrumentally observed data during 1781–1930 and 1940–1996 were separately analyzed. The partial regression coefficients of the AO and PDO in the multiple regression expression during 1781–1930 show that the AO and PDO are positively (0.96 ± 0.07 , $P < 0.00001$, $n = 150$) and negatively (-0.30 ± 0.06 , $P < 0.00001$, $n = 150$) correlated to summer RH with high significance, respectively (Table 2c). On the other hand, during 1940–1997, the partial regression coefficients of the AO and PDO in the multiple regression expression indicate that both of the AO and PDO are negatively correlated to the summer RH with high significances (-0.42 ± 0.15 , t test; $P < 0.00001$, $n = 57$ for AO and -0.41 ± 0.06 , t test; $P < 0.00001$, $n = 57$ for PDO) (Table 2c). This implies that the summer AO and annual PDO are independently related to the summer RH during 1781–1930 and 1940–1996.

[33] In Table 2c, the coefficient of determination explained by the summer AO ($R^2 = 0.56$) in single regression expression of the summer RH exceeds that explained by the annual PDO ($R^2 = 0.17$) before the 1930s. However, after the 1930s, the coefficient of determination obtained from the summer AO ($R^2 = 0.27$) in the single regression expression of the summer RH becomes less than that explained by the annual PDO index ($R^2 = 0.38$). The summer RH was dominantly regulated by the summer AO during 1781–1930 (the cold period), but it switched to the annual PDO during 1940–1997 (the warm period). By definition, the AO and PDO are related to the conditions of the Arctic climate and north Pacific climates, respectively. The climate transition from cold to warm periods due to global warming may change the cause of the variations in the hydrological climate in Hokkaido, northern Japan, from the forcing of the high latitude (AO) to that of the midlatitudes (PDO).

4. Conclusions

[34] The annual summer relative humidity (RH) since 1776 in Hokkaido, northern Japan, which is located in the monsoon pluvial area, were reconstructed using the oxygen isotope ratios in the tree ring cellulose of two living, oak trees. The result of MTM spectral analysis of the reconstructed summer RH indicated that the variation in the summer RH dominantly included the long-term periodicities of 50–100 years. We investigated the direct relationships between the decadal-centennial variations in the reconstructed summer RH and large-scale atmospheric circulation indices such as dendrochronologically reconstructed annual PDO indices [Biondi *et al.*, 2001; D'Arrigo *et al.*, 2001; MacDonald and Case, 2005] and summer AO index [D'Arrigo *et al.*, 2003] and the instrumentally observed

annual PDO and AO indices in order to understand the factors affecting the decadal-centennial changes in the hydrological climate in Hokkaido.

[35] The decadal-centennial variations in the summer RH are negatively correlated with those of the annual PDO index during all the periods considered. This is because the humid wind (southerly wind) from the western Pacific Ocean blows toward northern Japan with the intensified Pacific high and Asian continental low when the PDO index is low and vice versa. The decadal-centennial change in the summer hydroclimate in Hokkaido can be controlled by the subtropical North Pacific high and Asian continental low, which are related to the PDO.

[36] The decadal-centennial variations in the reconstructed summer RH are positively correlated with those of the reconstructed summer AO index during 1780–1930. On the other hand, the instrumentally observed summer RH during 1940–1997 is negatively correlated with the observed summer AO index on a decadal timescale. This drastic shift in its correlation is explained by the change in the atmospheric circulation during the 1930s due to the change from the negative to the positive phase of the AO index. The changed relationship between the decadal-centennial variations in the summer hydroclimate in Hokkaido and summer AO may be caused by structural changes in the AO pattern over the east Asian and Pacific regions.

[37] The summer RH is dominantly associated with the summer AO during 1781–1930 (the cold period) and annual PDO during 1940–1997 (the warm period). Global warming may dramatically change the causes of the variations in the hydrological climate in Hokkaido from the forcings of the high latitudes (AO) to those of the midlatitudes (PDO).

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