RECONSTRUCTION OF SUMMER TEMPERATURES IN INTERIOR ALASKA FROM TREE-RING PROXIES: EVIDENCE FOR CHANGING SYNOPTIC CLIMATE REGIMES

VALERIE A. BARBER [1], GLENN PATRICK JUDAY [1], BRUCE P. FINNEY [2] and MARTIN WILMKING [1]

[1] *Forest Sciences Department, University of Alaska Fairbanks, Fairbanks, AK 99775, U.S.A.*
E-mail: ffvab@uaf.edu
[2] *Institute of Marine Science, University of Alaska Fairbanks, Fairbanks, AK 99775, U.S.A.*

Abstract. Maximum latewood density and δ^{13}C discrimination of Interior Alaska white spruce were used to reconstruct summer (May through August) temperature at Fairbanks for the period 1800–1996, one of the first high-resolution reconstructions for this region. This combination of latewood density and δ^{13}C discrimination explains 59.9% of the variance in summer temperature during the period of record 1906–1996. The 200-yr. reconstruction is characterized by 7 decadal-scale regimes. Regime changes are indicated at 1816, 1834, 1879, 1916, 1937, and 1974, are abrupt, and appear to be the result of synoptic scale climate changes. The mean of summer temperature for the period of reconstruction (1800–1996) was 13.49 °C. During the period of instrument record (1903–1996) the mean of summer temperature was 13.31 °C for both the reconstruction and the recorded data. The coldest interval was 1916–1937 (12.62 °C) and the warmest was 1974–1996 (14.23 °C) for the recorded data. The reconstruction differs from records of northern hemisphere temperatures over this period, especially because of Interior Alaska warm periods reconstructed from 1834 to 1851 (14.24 °C) and from 1862 to 1879 (14.19 °C) and because of the cool period in the early part of the 20th century (1917–1974). We show additional tree ring data that support our reconstruction of these warm periods. Alternate hypotheses involving autogenic effect of tree growth on the site, altered tree sensitivity, or novel combinations of temperature and precipitation were explored and while they cannot be ruled out as contributors to the anomalously warm 19th century reconstruction, they were not supported by available data. White spruce radial growth is highly correlated with reconstructed summer temperature, and temperature appears to be a reliable index of carbon uptake in this system.

1. Introduction

Understanding how climate variability affects the functioning of ecosystems is of fundamental importance for natural resource management and ecological science. Recent studies of ice cores, lake-sediment, and tree-rings have documented many climate fluctuations during recent millennia. During this time decadal-to-century scale climatic changes were common, especially in the high latitudes, and included notable intervals such as the Medieval Warm Period (Hughes and Diaz, 1994) and the Little Ice Age (Bradley and Jones, 1993; Overpeck et al., 1997; Wiles and Calkin, 1990). Often these climate changes produce a long-lasting ecological imprint in basic ecological functions such as primary production or reproduction

Climatic Change **63:** 91–120, 2004.

of species. Occasionally a suite of proxy data from this climate-mediated pattern of ecosystem change can be assembled to reveal the fundamental pattern of climate variability itself.

A robust result of General Circulation Model (GCM) simulations is that high-latitude land masses in the northern hemisphere will experience the greatest magnitude of warming under scenarios of increased anthropogenically produced greenhouse gases (Houghton et al., 1996). There is growing paleoclimatic evidence that the 20th century was warmer than previous centuries (D'Arrigo and Jacoby, 1993; Jacoby et al., 2000, 1996; Mann et al., 1998). Equally impressive is the strength of evidence that the climate in certain high latitude regions has warmed markedly and abruptly in the last part of the 20th century (Chapman and Walsh, 1993; Houghton et al., 1996; Overpeck et al., 1997) and that large-scale ecological changes are already underway (Serreze et al., 2000). For example, strong warming since the 1970s in Alaska is associated with thawing permafrost (Osterkamp, 1996; Osterkamp and Romanovsky, 1999), receding glaciers (Wiles et al., 1996) and declining areal extent of Arctic sea ice (Chapman and Walsh, 1993; Serreze et al., 2000; Wadhams, 1995). Since the late 1970s, the area burned annually by wildfire in Canada has increased dramatically (Kasischke and Stocks, 2000; Kurz et al., 1995). During this same time period, insect outbreaks have killed trees over extensive areas in Alaska (Werner, 1996). It is tempting to interpret recent warming and ecological changes in Alaska as evidence for the greenhouse gas – climate-warming theory, but before doing so it is crucial to know whether other similar rapid climate changes and conditions occurred in the past. Paleo-records help define natural climate variability in order to assess potential anthropogenic changes in climate.

1.1. CONTROLS ON 20TH CENTURY CLIMATE

Interior Alaska is a well-defined, physiographically complex region bounded on the north by the Brooks Range and on the south by the Alaskan Range (63–67° N). Interior Alaska extends eastward to the Yukon Territory and westward to a climatic boundary (140–155° W) where precipitation exceeds 400 mm (Edwards et al., 2001). The region is made up of large, low-lying tectonic basins (Tanana Valley and Yukon Flats) separated by uplands (500–1000 m in elevation). The Brooks and Alaska Ranges act as topographic barriers to moisture-laden air from surrounding oceans. Consequently Interior Alaska is semi-arid, has a precipitation range of 400 to <200 mm annually (Patric and Black, 1968). Precipitation generally declines to the east, and is strongly influenced by topography (Edwards et al., 2001). About 60% of the annual precipitation falls as summer rain. The climate is cold continental with January mean temperature –20 °C or colder and July mean temperature 15–20 °C (depending on elevation and location within the region).

Unfortunately, climate records in Alaska are relatively sparse (most date from no earlier than mid-20th century). The oldest continuous record from Interior

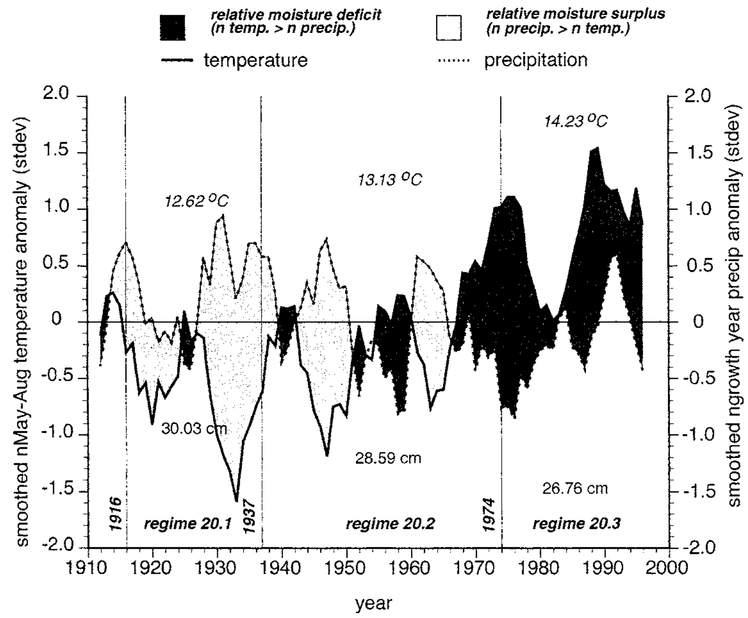

Figure 1. Growth year precipitation versus summer temperature anomalies (normalized data), from smoothed (5-yr running mean) Fairbanks recorded data. Boundaries of summer temperature regimes are indicated. Within-regime means are displayed for temperature and precipitation. Area of relative low summer temperature and relative high precipitation indicated as periods of moisture surplus; area of relative high summer temperature and relative low precipitation indicated as periods of moisture deficit.

Alaska is a combination of University Experiment Station (UES) and Fairbanks Airport data (Juday, 1984). The UES/Fairbanks record begins in the first decade of the 20th century. Fairbanks data are representative of the climate found in Interior Alaska. Average July temperature for Fairbanks is about 16 °C over the recorded period (1906–present), but has risen to about 17 °C when averaged over the last 20 years. Fairbanks annual precipitation is around 282 mm and has not increased over the past 20 years.

Interior Alaska has two distinct summer circulation patterns apparent in records of interannual climate. Twentieth century summer climate in Interior Alaska has alternated between periods of cool and moist or warm and dry conditions (Edwards et al., 2001; Mock et al., 1998; Juday, Barber, Rupp, Zasada and Wilmking, in press) (Figure 1). Summer climate is affected by mid-tropospheric variations of ridges and troughs with July and August normally the wettest months (Barry and Hare, 1974). When the Aleutian Low, the major synoptic climate feature of the North Pacific sector (Salmon, 1992), approaches the Alaska mainland from the southwest, precipitation can reach Interior Alaska unimpeded by topographic obstacles; all other avenues of approach cause precipitation to be intercepted on surrounding mountain slopes.

Warm and dry summer conditions in the Interior Alaska are caused by a high-pressure ridge located north to northeast of Alaska, which acts as a major circulation center. The high center brings clear skies and warm dry continental air from the east at the season of maximum surface heating from the long daylight hours near the Arctic Circle. Negative surface-pressure height anomalies over the Yukon Territory of Canada and over northern Siberia represent a northwest shift of the average pressure system. The Pacific subtropical high located south of Alaska is weak under this regime, resulting in reduced flow of moist air from the west. Persistent blocking ridge conditions are directly correlated with periods of extensive wildfires across the western North American boreal forest (Johnson et al., 1992).

The synoptic pattern for cool and moist conditions is produced by an eastward shift of the East Asian trough and a stronger-than-normal Pacific subtropical high. Both the eastward shift and the intensification of the subtropical high cause increased frequency of storms containing moisture-laden air to enter the interior basin from the southwest. The Brooks Range to the north and the Alaska Range to the south act as topographic barriers and prevent moisture-laden air from entering from these directions.

Circulation patterns associated with; (1) cold and dry or (2) warm and wet summer conditions in Interior Alaska are rare. For example, recent studies (Edwards et al., 2001; Mock et al., 1998) showed there was only one year each of anomalously warm/wet and cold/dry July climate between 1946–1989. Cold and dry conditions appear to result from a southward displacement of the jet stream as compared to normal with a low-pressure center and a westerly wind maximum far south of Alaska into British Columbia (Mock et al., 1998). North of this low-pressure center, colder and drier air masses predominate, a condition that resembles spring circulation. Warm and wet conditions set up when low-pressure centers are centered more westward than normal and a high-pressure center is prevalent over Alaska. Convective activity might explain the positive precipitation anomalies during the time that temperature anomalies are high (Mock et al., 1998).

1.2. ECOLOGY OF WHITE SPRUCE AND ITS CLIMATE SENSITIVITY

White spruce occupies a well-defined climate distribution of July mean temperature from about 8 °C to 21 °C, with total July precipitation between about 20 mm and 200 mm, as indicated by; (a) response functions generated by modern pollen studies across the distributional limits of the genus *Picea* (Anderson et al., 1991) and (b) the relationship of climate with the modern distribution of *Picea glauca* (Thompson et al., 2000). Actual or realized climatic limits of the species are defined by the interaction of summer temperature and precipitation (Nienstaedt and Zasada, 1990).

In western Canada, the northern extent of white spruce appears to be limited by lack of warmth while the southern extent is controlled by lack of moisture (Brooks et al., 1998). At the southern margin of the Canadian boreal forest in the Prairie

Provinces of central and western Canada, forest distribution is limited to areas in which the difference between precipitation and potential evapotranspiration is positive (Hogg, 1994, 1997). Forest distribution in the southern boreal region of western Canada is controlled by chronic moisture deficits (Hogg, 1997).

In Interior Alaska, precipitation is near the limit for occurrence of the species and upland white spruce on low-elevation sites occupies both the warm and dry margin of the climatic range for the species (Thompson et al., 2000). Potential evapotranspiration during the growing season is often greater than annual precipitation (Patric and Black, 1968). In such environments, evapotranspiration as influenced primarily by high summer temperature has been interpreted as the limiting factor in annual radial growth (Barber et al., 2000). A negative relationship between summer temperature and radial growth of white spruce on warm, low elevation productive sites in Interior Alaska is consistent throughout the entire 20th century (Barber et al., 2000).

Most dendrochronological literature on white spruce (*Picea glauca* (Moench) Voss) in western North America is based on treeline collections (Garfinkle and Brubaker, 1980; Jacoby and D'Arrigo, 1989). In such studies, sampled trees are assumed to be growth limited by a single climate variable (usually temperature) and the trees are free from canopy competition. The literature is dominated by studies of trees displaying a positive ring-width response to summer temperature (Jacoby and D'Arrigo, 1989). However, in the second half of the 20th century, trees at treeline across high northern latitudes have become less sensitive to temperature at many locations. In such treeline trees, additional summer warmth has produced either limited or no additional growth (Briffa et al., 1998), and lack of moisture may now be limiting growth in contrast to low temperatures (Jacoby and D'Arrigo, 1995; Lloyd and Fastie, 2002).

A much greater area and the majority of biomass production of white spruce forest occur in lower elevation stands (Ruess et al., 1996) rather than marginal treeline stands. White spruce-dominated forest types make up about 12 million ha or 26% of the Alaska boreal forest, including 2.8 million ha or 51% of the commercially productive forest area (Labau and Van Hees, 1990). Thus changes in the annual growth of low-elevation, productive white spruce forest types will be a major factor controlling variability of carbon dioxide uptake in the boreal forest of western North America.

In this paper we examine ring-width, maximum late-wood density and δ^{13}C discrimination of annual tree-rings of white spruce from Interior Alaska sites (Figure 2) to reconstruct and interpret 19th century climate. We explore the relationship of these tree-ring properties to temperature, precipitation and potential evapotranspiration. Both δ^{13}C discrimination and maximum latewood density of white spruce tree-rings are highly correlated with Fairbanks summer temperature during the 20th century (Barber et al., 2000) and δ^{13}C discrimination represents a physiological signal of drought stress internal to the tree (Francey and Farquhar, 1982). We use the relationship of combined (equally weighted) δ^{13}C discrimination and maximum latewood density to mean summer temperature (May–August) in a principal components regression analysis (using the program PCREG) to model summer temperature in the 19th century when climate records are not available. We also evaluate radial growth of white spruce and other tree species in Alaska to determine the consistency of tree growth responses to reconstructed summer temperature.

2. Methods

All tree-rings in the sample used for our analysis are representative of white spruce in mature and old stands that are dominant on the contemporary landscape of Interior Alaska, including trees across a broad range of diameters. We measured three properties of tree-rings, stable carbon isotope ratios ($\partial^{13}C$), maximum latewood density and ring-width. Trees at each site were crossdated with the software Cofecha (Holmes, 2000) and were visually inspected for marker years to insure year-to-year correspondence in all measurements.

For development of ring-width chronologies, we used the 10 oldest stands across Interior Alaska from the 20-stand calibration set used in Barber et al. (2000) (Figure 2). Trees at each individual site were crossdated with Cofecha and individual trees were corrected where possible if a problem was identified. Trees that were not significantly correlated with the mean chronology were excluded. Sample depth for individual trees during the period 1800 to 1996 varied between 43 and 220 trees (Figure 3). Ring-widths were standardized with the program ARSTAN (Holmes, 2000). Standardization removes the age-related trend by fitting a curve to each tree-ring series. This removes most of the low frequency variation, gives dimensionless indices and prevents faster growing trees from dominating the interannual variability. We used conservative negative exponential or straight-line curve fits (Fritts, 1976) and a master chronology was created from the individual trees for each site. A master chronology for Interior Alaska was then created using the 10 individual stand chronologies as the ARSTAN input. Sample depth of stand number varied between 2 and 10 (Figure 3) during the period of analysis (1800–1996). Ring-width chronology values prior to 1816 had fewer than half of the stands or individual trees contributing to the chronology and relationships between growth and reconstructed climate in those years should be treated with caution because of low sample depth (Figure 3). We also used ring-width chronologies of trees from a site in south-central Alaska (Lutz spruce), a site in the Brooks Range (white spruce), and another site in Bonanza Creek LTER (black spruce). The same crossdating and standardization were employed as above. Sample depth was more limited at these sites, but these were used primarily for detection and corroboration of growth responses consistent with our temperature reconstruction.

Stable carbon isotope ratios were measured only at the Bonanza Creek LTER Reserve West stand. Growth patterns of dominant trees at Reserve West were

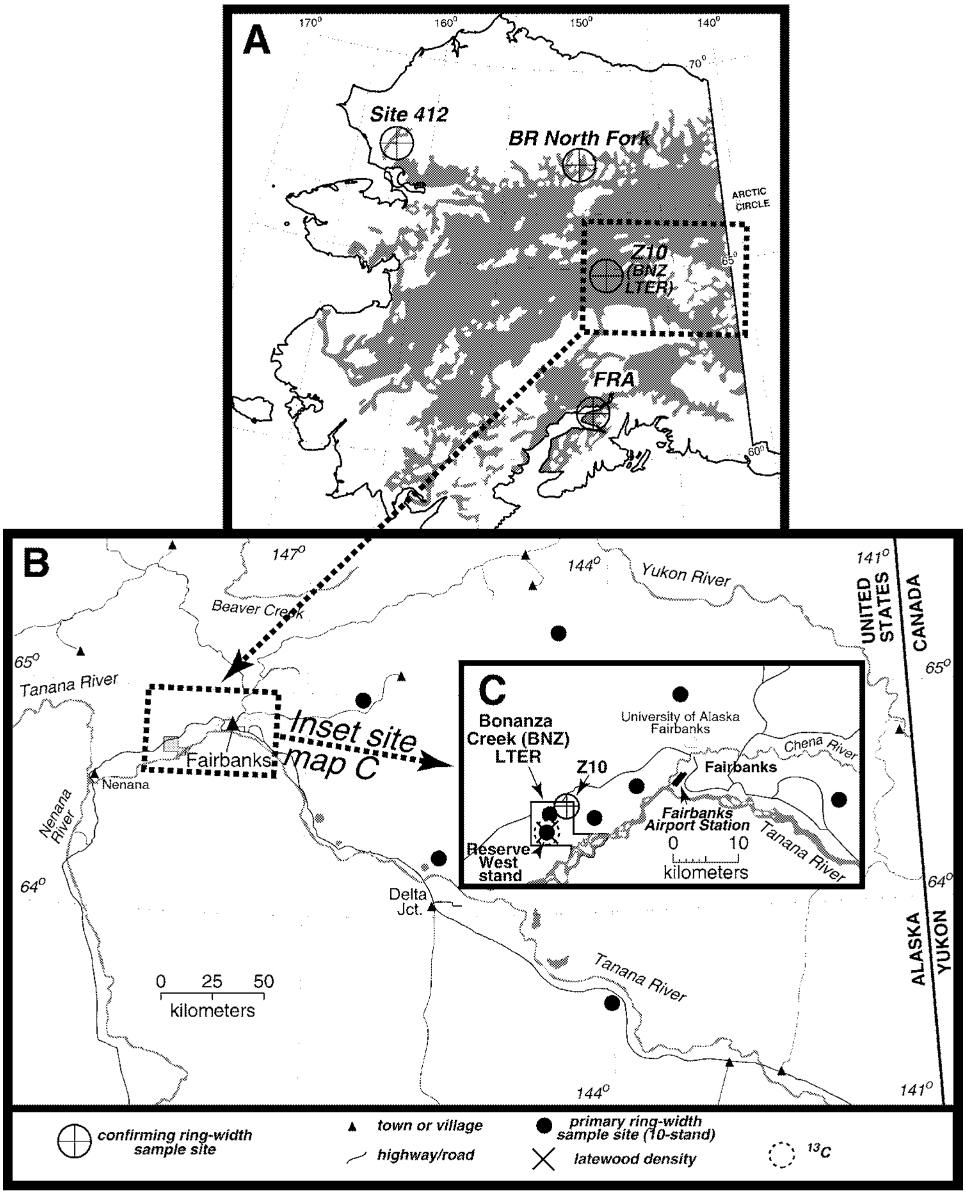

Figure 2. Location of tree-ring sampling sites in interior Alaska. A: Map of Alaska with boreal forest shaded. Symbols at location of additional corroborating tree-ring sample sites; Site 412 at northwest Alaska tree limit, BR North Fork at central Brooks Range treeline, Z10 at Zasada Road Site 10 in Bonanza Creek Long-Term Ecological Research (BNZ LTER) site, and FRA at Fort Richardson in Anchorage in southcentral Alaska. B: Expanded view of central Alaska with outlying sample sites of 10-stand upland white spruce sample. C: Expanded view of Fairbanks vicinity including BNZ LTER and Fairbanks International Airport climate station. Note the co-occurrence of ring-width, latewood density, and $\delta^{13}C$ sampling in the Reserve West stand at Bonanza Creek Long-Term Ecological Research (LTER) site, and the proximity of the Z10 black spruce sample site.

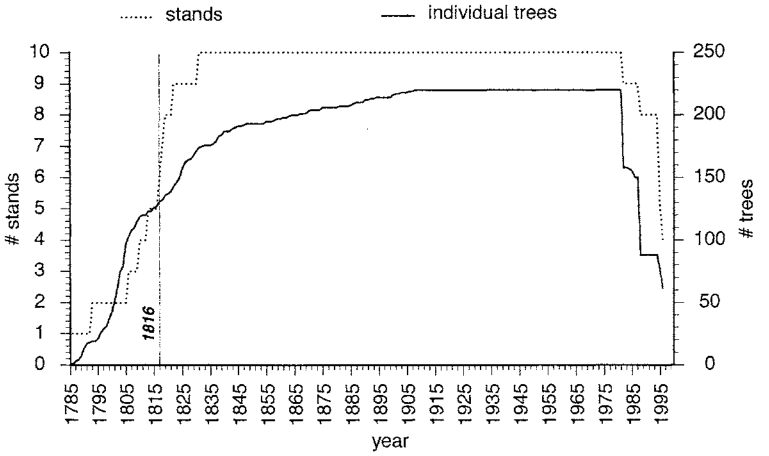

Figure 3. Sample depth for 10-stand upland white spruce ring-width sample calculated as number of trees and number of stands. Note the fall-off in sample depth before 1816.

highly correlated with the other 9 stands and the Fairbanks summer temperature record, suggesting the existence of a common signal. Four orthogonal cores from each of 4 trees were pooled together to give an isotope value representative of the site (Leavitt and Danzer, 1992). We obtained isotope measurements for the period 1800–1981 from 4 orthogonal wedges from each of 4 harvested stump disks of trees killed in a 1983-forest fire. For the period 1967–1996, isotope samples were obtained from 4 cores collected at each of 4 surviving trees near the fire perimeter. The surviving trees were the same age class as burned trees; projected date of stand origin was 1785 based on a charcoal layer in the soil and a ring count of stump sections. Burned and surviving trees occurred at the same elevation (290 m asl) and aspect (azimuth 150°), and grew on the same soil type (Fairbanks silt loam – Alfic cryochrept in loess parent material). (Details of the site description can be found at http://www.lter.uaf.edu/Site_detail.cfm.)

Four orthogonal samples of annual rings were excised, ground, and blended for each tree and then combined with the same-year wood from the other trees. Hollocellulose was extracted from the annual wood (Leavitt and Danzer, 1992) and carbon isotope ratios were analyzed on a Europa 20/20 continuous flow mass spectrometer for the years 1800–1996. For some years, hollocellulose carbon isotope ratios were measured for individual trees and differences were sometimes as much as 1.0‰, but trends were consistent throughout the 200 years measured and all trees were significantly correlated (0.6–0.75). Isotope values are reported relative to the PDB standard and the precision of the analyses is $\pm 0.1‰$. The overlap period, 1967–1981, was used to determine site-specific isotopic offsets between the fire-killed and surviving trees. Isotope trends for the period of 1967–1981 follow similar curves from both sets of wood samples, but there is a slight offset (0.7‰) in the $\delta^{13}C$ of the fire-killed vs. the live trees likely due to site-specific effects. A

Table I

Statistics on the correlation of May–Aug. temperature at Fairbanks (1906–1996) with standardized maximum latewood density at three individual sites. *N* refers to the sample size and Total # years refers to the age of the oldest trees

	r	*P*	*N* (1906–1996)	Total # years
Reserve West	0.707	0.00001	13	215
Dune Lake	0.649	0.00001	4–18	103
Jan Lake	0.547	0.00001	5–17	190 (one core)

correction was made by adding this offset to the yearly isotope values of the 1982–1996 data set to bring it into the same range as the longer data set. To correct for δ^{13}C changes in atmospheric CO_2 over the last 150 years due to fossil fuel and combustion, we used the Law Dome Antarctic ice core data (1.53‰change over 150 years (Francey et al., 1999)). Discrimination was calculated as the difference between the δ^{13}C of the atmosphere (ice core) and the measure of δ^{13}C in the wood. Thus a continuous record of δ^{13}C discrimination was determined for 1800–1996. We confirmed the trends in δ^{13}C seen at Reserve West in limited samples covering 3 to 4 decades collected from 2 other sites located within 30 km of the Reserve West site.

Maximum latewood density was measured by x-ray attenuation at Lamont-Doherty Earth Observatory. Tree cores were collected from 3 of the 10 sites (Dune Lake (64°25′ N, 149°54′ W), Jan Lake (63°34′ N, 143°54′ W) and the Reserve West stand at Bonanza Creek LTER (64°44′ N, 148°18′ W) for density analysis. A total of 14 cores from 12 trees were used from Reserve West, 19 cores from 9 trees for Jan Lake and 18 cores from 8 trees from Dune Lake. Density chronologies for individual trees from each site were run separately through ARSTAN and a master standardized chronology was created for each site. Unfortunately, all but one of the Dune and Jan Lake trees were under 100 years old (Table I), not long enough for the reconstruction of the 19th century, or lacked sufficient sample depth (Table I). Since all three sites were highly correlated with each other (Table II) and with the Fairbanks climate, only the Reserve West maximum latewood density chronology, which was over 200 years in length and of sufficient sample depth, was used for the reconstruction of Fairbanks summer temperatures.

We first established which climate variables had significant relationships with the standardized chronologies of ring-width, δ^{13}C discrimination, and maximum density. We determined relationships of monthly temperature and precipitation to the standardized chronologies for growth year (Sept.–Aug.) and for year prior to growth (lag –1) using the principal components multiple regression analysis (Cook and Kairiukstis, 1990). Significant relationships were restricted to April

Table II

Pearson correlation coefficients between the three maximum latewood density sites

	Reserve West	Dune Lake	Jan Lake
Reserve West	1.0	–	–
Dune Lake	0.567	1.0	–
Jan Lake	0.646	0.740	1.0

through August temperature, but May contained essentially all the information in April. We determined that the best relationship between the chronologies and climate involved mean May through August temperature. All three tree-ring parameters, δ^{13}C discrimination, maximum latewood density and 10-stand ring-width, are highly correlated with mean summer (May–Aug.) temperature at Fairbanks for the contemporary growth year (Barber et al., 2000). However, ring-width is the most autoregressive tree-ring parameter because it is significantly correlated with summer temperature one and two years prior to the year in which the ring was formed (Barber et al., 2000). Maximum latewood density and δ^{13}C discrimination are much less autocorrelated and are most highly correlated with summer temperature in the year of ring formation (Barber et al., 2000).

The significance of the relationship between each tree-ring parameter and summer temperature was tested in a calibration and verification step over the period of instrument record (1906–1996). A reconstruction was produced based on data from the first half of the period of record (calibration step) and then the significance of the reconstruction during the second half of the period (verification step) was evaluated. The calibration-verification process was repeated in reverse. Only tree-ring parameters that passed both tests were use in the final reconstruction over the entire 1800–1996 time period.

Since the reconstruction showed distinctive decadal to multi-decadal periods, we divided it into regimes. In order to establish boundaries of regimes, we used moving split window analysis with squared Euclidean distance metrics (MW SED) (Johnson et al., 1992; Turner et al., 1991) on the May–August temperature reconstruction. Unprocessed reconstructed summer temperature data (1800–1996) subjected to spectrum analysis contains intermediate peaks at 9.1 and 18.2 years. However, a preprocessing (differenced, 64-year segments sampled at 32-year intervals) treatment of the 196-year reconstructed temperature showed no significant periods at the 95% confidence level. We chose a 34-year period for the moving window because half of it is one of the two odd-integer terms (17 and 19 years) that brackets the longest quasi-cycle (18.2 years). This approach compares the average of 17 years with an average of the following successive 17 years and optimizes the definition of regimes by maximizing index values at natural periods of change. Empirical investigation of a series of potential odd integer terms (11, 13, 15, 19, 21) demonstrated broad similarities in identified regimes (periods between successive change index maxima).

3. Results

3.1. THE 200-YEAR RECONSTRUCTION

The best reconstruction of summer (May through August) temperature by a single tree-ring parameter was produced by δ^{13}C discrimination, which explained 46.3% of the variance (Table III) during the period of the temperature record at Fairbanks. Both the calibration and verification periods produced a significant correlation (Table III). Maximum latewood density also produced a summer temperature reconstruction with high, but slightly lower statistical significance, explaining 38.7% of the variance (Table IV). The reconstruction of summer temperature produced by ring-width alone was significant over the entire period of instrument record, explaining 29.6% of the overall variance, but it was not significant during the first half of the calibration period (1906–1950) (Table V), thus we excluded ring-width from our final reconstruction of summer temperature.

The final reconstruction based on combined δ^{13}C discrimination and maximum latewood density (Figure 4) explained more of the overall variance (59.9%) and more of the variance in the calibration and verification periods than any of the individual ring parameters (Table VI). Adding ring-width to the reconstruction did not increase the explained variance (in fact it declined slightly) and thus we used only δ^{13}C discrimination and maximum latewood density for the final reconstruction. For the earlier calibration period (1906–1950) of the final reconstruction, the adjusted r^2 is 0.499. The RE (reduction of error) statistic is strongly positive (0.710) over the verification period (1951–1996), which shows that there is considerable skill in the verification estimates as compared to the calibration period mean. However, it may also partly reflect a difference between the means of the calibration and verification periods. The coefficient of efficiency (CE) statistic differs from the RE in that it compares the estimated data for the verification period to the mean of this period and this number is also strongly positive for both halves of the calibration/verification data. For additional verification, the Spearman rank correlation and product means test statistics were calculated and both were significant at the 0.0001 level. The cross-product means test measures the level of agreement between the actual and estimated values and takes into account the sign and magnitude of departures from the calibration average (Fritts, 1976). When the calibration and verification periods are reversed, the adjusted r^2 is 0.511, the RE is 0.710, the CE is 0.408, the Spearman rank correlation is 0.727, and the cross product mean test is 0.499, all highly significant. These results indicate that the

Table III

Calibration-verification statistics for reconstruction of May–Aug. temperature at Fairbanks based on ^{13}C discrimination

Calibration ^{13}C discrimination	1906–1996	1906–1950	1951–1996
r	–0.680 ^{c}	–0.576 ^{c}	–0.648 ^{c}
r^2	0.463	0.332	0.420
Adj. r^2	0.457	0.317	0.407
S		0.659 ^{c}	0.546 ^{c}
RE		0.569	0.573
CE		0.141	0.128
Sign test		29+ 17–	28+ 17–
Cross product means test		0.443 ^{b}	0.483 ^{b}

r = multiple correlation coefficient for predictor 1 (^{13}C discrim.).
r^2 = variance explained.
Adj. $r^2 = r^2$ adjusted for loss of degrees of freedom.
S = Spearman rank correlation coefficient.
RE = reduction of error statistic.
CE = coefficient of efficiency.
[a] Significant at the <0.01 level.
[b] Significant at the <0.001 level.
[c] Significant at the <0.0001 level.

Table IV

Calibration-verification statistics for reconstruction of May–Aug. temperature at Fairbanks based on maximum latewood density

Calibration model max. lw. density	1906–1996	1906–1950	1951–1996
r	0.622 ^{c}	0.540 ^{c}	0.565 ^{c}
r^2	0.387	0.292	0.319
Adj. r^2	0.380	0.275	0.304
S		0.547 ^{c}	0.531 ^{b}
RE		0.494	0.477
CE		–0.008	–0.082
Sign test		27+ 19–	29+ 15–
Cross product means test		0.439 ^{c}	0.208

Symbols as on Table III.

Table V

Calibration-verification statistics for reconstruction of May–Aug. temperature at Fairbanks based on ring-width

Calibration model ring width	1906–1996	1906–1950	1951–1996
r	–0.544 c	–0.299 [a]	–0.548 [c]
r^2	0.296	0.090	0.301
Adj. r^2	0.288	0.068	0.285
S		0.495 [b]	0.299 [a]
RE		0.367	0.435
CE		–0.263	–0.155
Sign test		21+ 25–	28+ 17–
Product means test		0.355	0.171

Symbols as on Table III.

Table VI

Calibration-verification statistics for reconstruction of May–Aug. temperature at Fairbanks based on maximum latewood density and ^{13}C discrimination

Calibration model ^{13}C dis and MLWD	1906–1996	1906–1950	1951–1996
r_1	–0.680 [c]	–0.576 [c]	–0.648 [c]
r_2	0.622 [c]	0.540 [c]	0.566 [c]
r^2	0.599	0.511	0.522
Adj. r^2	0.595	0.499	0.511
S		0.719 [c]	0.727 [a]
RE		0.722	0.710
CE		0.445	0.408
Sign test		38+ 8–	33+ 12–
Cross product means test		0.543 [c]	0.499 [c]

Symbols as on Table III, with
r_1 = multiple correlation coefficient for predictor 1 (^{13}C discrim.).
r_2 = multiple correlation coefficient for predictor 2 (max dens.).

model used here (combined δ^{13}C discrimination and maximum latewood density) passes the critical tests for verification and is optimized by achieving the greatest predictive capability for the fewest independent variables.

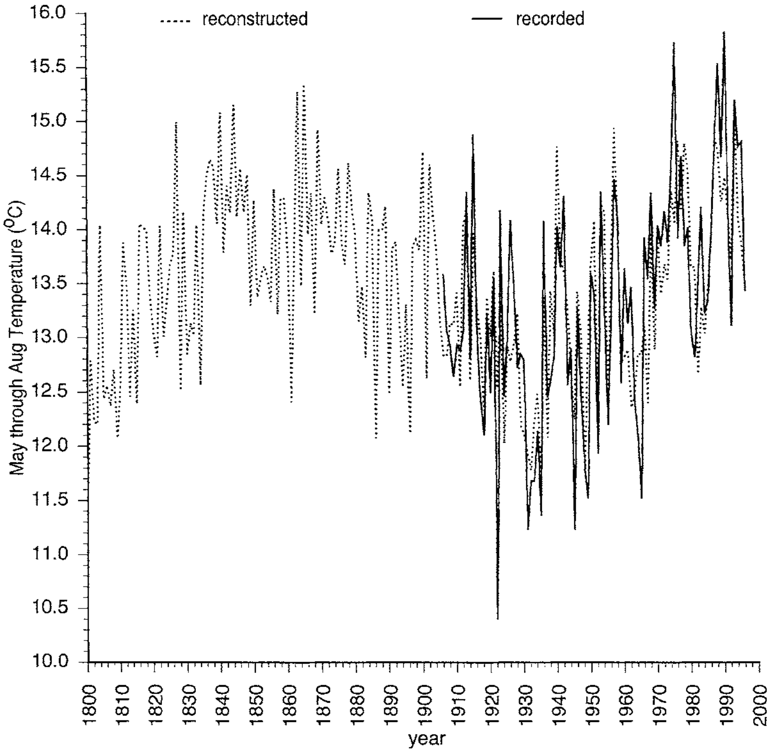

Figure 4. Reconstruction of May–August temperature for Fairbanks Alaska based on ^{13}C discrimination and maximum latewood density compared to recorded data.

The average reconstructed summer temperature for the entire period (1800–1996) is 13.49 °C (Figure 5). The mean of both the 20th century (1906–1996) recorded temperature and the reconstructed temperature is 13.31 °C.

3.2. IDENTIFICATION OF SUMMER TEMPERATURE REGIMES

The temperature reconstruction for the 200-year time period shows a distinctive low frequency sinusoidal pattern with high-resolution decadal-scale periods overlain (Figure 4). These decadal-scale periods shift rapidly over the course of a few years.

The results of the MW SED show peaks or spikes where the greatest change occurred, and we used these spikes to define the boundaries of climate regimes. Our proposed climate regimes are defined as multi-decadal periods of characteristic summer mean temperature separated by periods of rapid climate changes of a few years in duration (Figure 5). For convenience, we number climate regimes for the century in which they began, and assign a decimal of 1 for the first regime initiated in the century and increasing decimal numbers for successive regimes of that century. Thus the first regime initiated in the 20th century is labeled 20.1, etc.

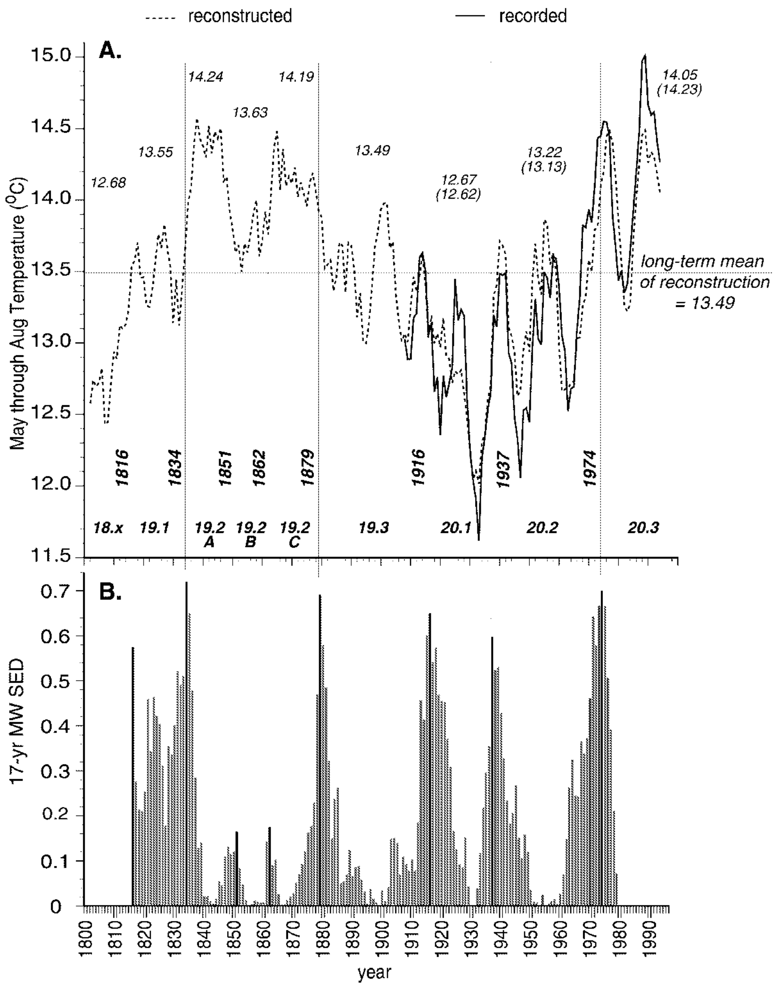

Figure 5. A. Regimes of summer temperature including dates of change. Note identifying regime numbering system composed of century identifier (left of decimal) and sequential numeral within the century (right of decimal) along horizontal axis. Reconstructed mean temperature is displayed for each regime (recorded mean in parentheses). B. 34-year moving split-window squared Euclidean distance (MW SED) metric used to define nodes of change.

We have further divided one regime (19.2) into three sub-regimes (A, B, and C) of lesser magnitude change.

Based on these criteria, the 20th century contains Regimes 20.1, 20.2, and 20.3, and the 19th century is divided into 3 regimes. We tentatively identify a rapid climate change at about 1816 based on an apparent major change in variables compared to Regime 19.1, although we do not have MW SED values calculated prior to that year. These early years of the 19th century appear to have been part of a regime that began in the 18th century and since we are unsure of the number of regimes contained in that century, we label that regime as 18.x.

3.3. PRECIPITATION AND EFFECTIVE MOISTURE

Adding a precipitation term to a synthetic climatic index that predicts radial growth of white spruce in Interior Alaska can improve the correlation score over a climate index made up of temperature terms alone (Barber et al., 2000). We also examined the relationship of δ^{13}C discrimination and maximum latewood density to monthly precipitation and did obtain significant results for July and August. We combined July and August precipitation and the principle components regression model results showed that δ^{13}C discrimination had a positive relationship to precipitation ($r = 0.318$) while maximum latewood density had a negative relationship ($r = -0.469$), supporting our moisture stress hypothesis. The variance explained by the regression model results using these two proxies to reconstruct July plus August precipitation was only 21.7%, much less than the 59.9% explained when summer temperature alone was reconstructed by δ^{13}C discrimination and maximum latewood density.

In an effort to look more closely at the effects of moisture, we examined the relationship of effective moisture (precipitation minus potential evapotranspiration (PET)) to the tree ring proxy data. The calculated PET was based on monthly minimum, maximum and mean temperature (May–Sep) (Hogg, 1997) and precipitation was calculated as accumulated growth year precipitation (Oct–Sept). While this is a simplified method for calculating effective moisture, it does produce some interesting and relevant results. We looked at simple correlation coefficients of effective moisture to δ^{13}C discrimination ($r = 0.45$), maximum latewood density ($r = -0.27$) and ring width ($r = -0.36$). The fact that effective moisture is positively related to δ^{13}C discrimination and negatively related to maximum latewood density supports our interpretation that the mechanism by which summer temperature controls these tree-ring properties is temperature-induced drought stress. However, the relationships are not as significant as temperature alone and so the most successful reconstruction of climate that can be obtained from our tree-ring proxy data is summer temperature.

4. Discussion

4.1. INTERPRETATION OF THE RECONSTRUCTION

The 200-year reconstruction based on the combination of tree-ring proxies δ^{13}C discrimination and maximum latewood density, shows excellent agreement with the recorded Fairbanks mean May through August temperatures (Figure 4). Recorded annual summer temperature displays greater amplitude than reconstructed temperature, which indicates that although δ^{13}C and density are superior to ring-width in reconstructing summer temperatures on an annual basis, these parameters still retain a small amount of autoregression as seen by the correlation with prior-year July temperature (Barber et al., 2000). Some loss of variance and amplitude is also expected in a regression due to the unexplained variance. Both the reconstructed and recorded summer temperatures of the latter part of the 20th century, particularly from 1970 onward, are characterized by some of the warmest summers in the 200-year interval (Figure 5). The first half of the 20th century is characterized by the coolest summers of the 200-year period of reconstruction. Interestingly, mid-19th century summer temperatures reconstruct as some of the warmest over the 200-year period.

In an attempt to look more closely at our reconstruction of summer warmth during the mid-19th century, we looked for differences in the individual reconstructions of summer temperatures by the three proxies (δ^{13}C discrimination, maximum latewood density and ring-width index) (Figure 6). All three proxies by themselves reconstruct overall warmer summer temperatures in the 19th century than temperatures in the early to mid-20th century. There are some discrepancies between the three proxies, especially for the early part of the 19th century (1800–1815). The δ^{13}C discrimination-based reconstruction shows the greatest anomaly and indicates cooler temperatures in the earliest decade of the 19th century than the other two proxies do. Carbon isotopes also reconstruct much warmer temperatures from the mid-1850s to 1879. Isotopes and density reconstruct increasing temperatures around 1835 while ring width indicates cooler temperatures. These discrepancies could be artifacts of sampling and/or they could reflect ecological differences due to micro-site conditions. We would expect more differences between ring width and the other proxies because the ring-width sample is made up from 10 different sites throughout interior Alaska while the δ^{13}C discrimination and maximum latewood density are from trees from the same site, although not necessarily from the same trees. In hindsight, sampling strategy would have been slightly different based on knowledge gained from this project. As it was, data analysis was biased toward availability of samples. However, compared to regimes 20.1 and 20.2 (1916–1974), all proxies show either relatively warm or moderate temperatures from about 1860 through 1879, especially when compared with the early part of the 20th century. The density and δ^{13}C discrimination records show good agreement and higher temperatures from about 1830–1879. Thus, all three proxies agree on the interpretation of a fairly warm mid-19th century followed by cooling into the middle decades of the 20th century.

Most reconstructions of Northern Hemisphere temperature indicate low annual temperatures in the earlier part of the 19th century, which is generally attributed to the latter part of the Little Ice Age (Bradley and Jones, 1993). Some published reconstructions for the Northern Hemisphere indicate a warming in the mid-19th century that continued into the early 20th century (Jacoby and D'Arrigo, 1989; Overpeck et al., 1997). In contrast our data show that the coolest part of the 200-year reconstructed summer temperature for Interior Alaska was the early part of the

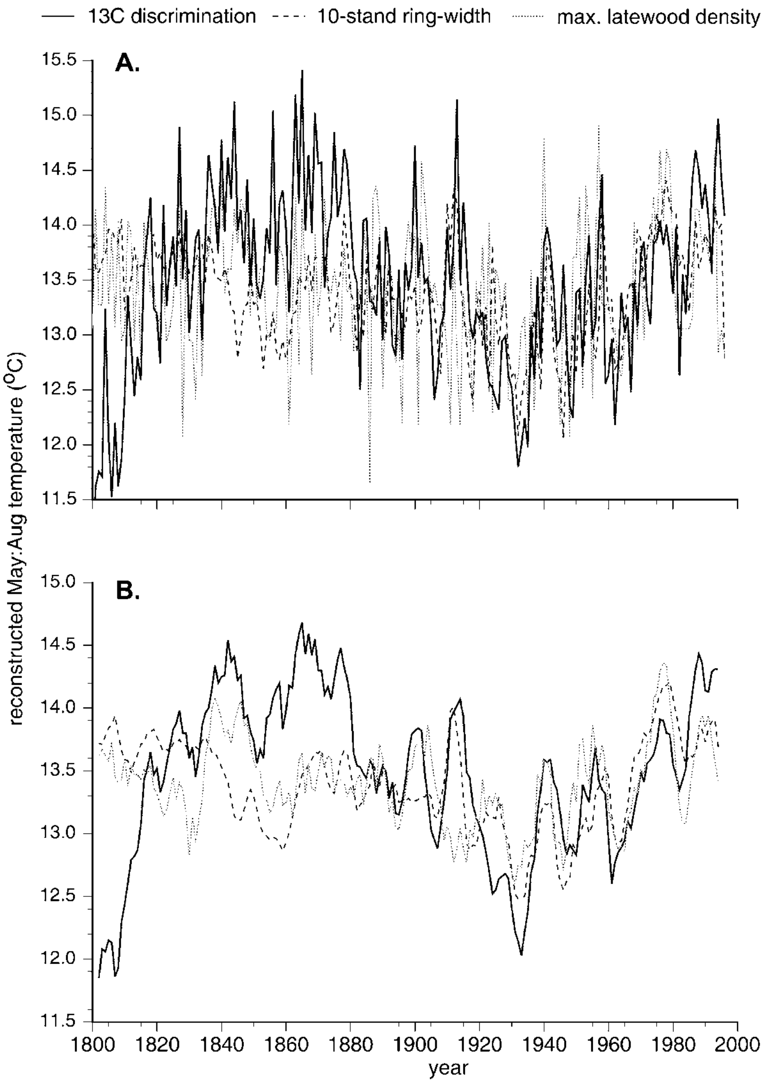

Figure 6. Summer temperature reconstruction at Fairbanks based on three individual proxies of ^{13}C discrimination, maximum latewood density, and ring-width index. A. Annual values. B. Smoothed (5-yr. running mean) values.

20th century (1916–1937), a cool interval not seen in most Northern Hemisphere temperature reconstructions or records (Jacoby and D'Arrigo, 1989; Mann et al., 1998; Overpeck et al., 1997). Our results suggest that in fact, Interior Alaska probably experienced temperature trends different from the overall trend in northern North America during the period of analysis. For example, Northern Hemisphere mean temperatures (Mann et al., 1998) (Figure 7) show a fairly steady increase in warm season temperatures anomalies from the earliest 20th century until the

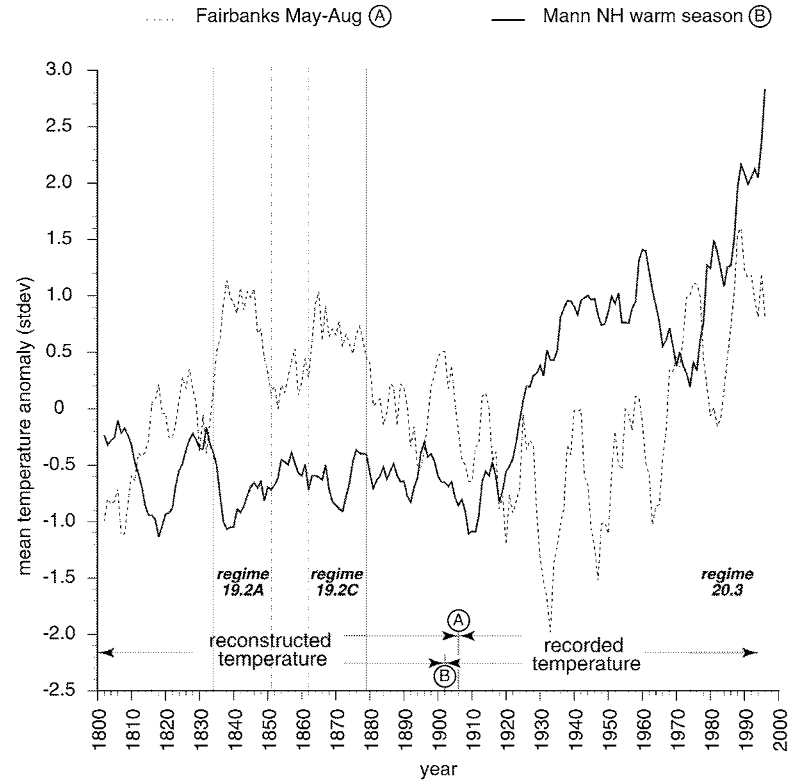

Figure 7. Smoothed anomalies of Fairbanks mean summer temperature vs. Northern Hemisphere warm season temperature (Mann et al., 1998). Note inverse relationship on quasi-decadal scale, including during period of recorded data and reconstructed warm periods of Regime 19.2A and 19.2C.

mid-1950s, while Fairbanks recorded mean summer temperature shows a cooling from around 1916 through 1937 (regime 20.1). The Fairbanks warm season record displays a brief warming in the early 1940s and at 1957–58. Summer temperatures remain generally cool from 1958 until the mid 1960s, after which they increase to the present. The Northern Hemisphere warm season temperature anomaly shows little trend from the mid-1950s to about the mid-1980s at which point the two records show the same trend of warming to the highest levels of the century (Figure 7).

There are notable differences between the two records and there appears to be an opposite trend in temperature during some regimes. There is also much greater variability and larger amplitude in the Fairbanks record than in the Mann et al. (1998) record and reconstruction. Differences in the two records suggest that regional controls probably due to large-scale atmospheric circulation patterns can have different impacts on Alaska relative to the rest of North America, such as during the earlier part of the 20th century. The common trend of rising temperature during the latter part of the 20th century is consistent with common forcing from greenhouse gases although there are still some contrasting patterns. However, the warming in the latter part of the 20th century in Alaska is clearly more than can be directly attributed to external forcing from factors such as CO_2 and volcanic eruptions. The discrepancies between the two temperature records, and the large amplitude changes in the Interior Alaska summer temperature record and reconstruction, clearly indicate the importance of regional-scale atmospheric circulation in determining the climate of this region. Such patterns may have occurred in the past, and could account for the relatively warm summer temperatures we reconstruct during the mid-1800s.

Further evidence is available from gridded temperature patterns in the Northern Hemisphere (Mann et al., 1998) based on multiproxy EOF reconstructions that reveal anomalies in Alaska temperatures as compared with the rest of northern North America. One interesting result from that study shows that most of northern North America was very cold around 1834, but Alaska was warmer than average, which is consistent with our reconstruction. An historical account from 1822 given by Nordkvist from the Bering Sea region hints at relatively warm conditions in Alaska in 1822; ‘Not only in the summer, but in the winter the ocean was free of ice sometimes with a wide strip of water up to at least 200 miles away from the shore’ (Koskey and Yamin, 2001). Thus our reconstruction may be due to a large-scale atmospheric circulation pattern that resulted in a localized regionally warmer climate for Interior and northern Alaska, but a cooler climate elsewhere over much of northern North America.

4.2. ALTERNATIVE HYPOTHESES OF 19TH CENTURY CLIMATE

Two alternative hypotheses could account for our reconstruction of warm temperatures in the middle of the 19th century:

1. The proxy signal (δ^{13}C discrimination and maximum latewood density) upon which the reconstruction is based could have been influenced by properties of the site or trees that were unique to the early life of the stand (juvenile effect), a time that happens to correspond to the early to mid-19th century. Once the trees or stand matured, these effects disappeared and the period of reconstruction overlap with recorded climate data essentially represents a different calibration interval.
2. Conditions with no modern analog could have existed in the 19th century. There are 2 parts to this hypothesis:

   (a) It could have been colder than the range of temperature seen in the 20th century and the trees may have been limited by the direct effect of very low temperatures. This explanation would require that an ecological threshold be crossed so that the trees used for the reconstruction in this paper and on similar sites in Interior Alaska, changed their growth response to temperature from positive (19th century) to negative (20th century).

(b) It could have been moderately cool to cold but very much drier in the 19th century. There are no sustained periods of cold and dry summer conditions in the 20th century record as indicated earlier in this paper. We demonstrated that drought stress induced by high temperatures was the limiting factor behind reduced radial growth in white spruce on productive sites in central Alaska in the 20th century (Barber et al., 2000). According to this interpretation, extremely arid conditions in the 19th century could have produced physiological drought in the trees directly (without the influence of high temperatures) and thereby limited their growth, even during a period of cooler summer temperatures than during the 20th century.

With regard to hypothesis 1, there could be a juvenile effect in the isotope data (Bert et al., 1997), which result in discrimination being driven down as the trees grow, the canopy expands, and evapotranspiration increases. Autogenic effects on a site are prominent early in the life of a forest and are known to influence moisture stress (Bert et al., 1997). The trees used for δ^{13}C analysis came from a stand that regenerated around 1785. As the trees grew in the first several decades of the life of the stand, there may have been an increase in moisture stress due primarily to the development of a full canopy. Canopy expansion would thus produce a trend toward greater evapotranspiration and less δ^{13}C discrimination. However, such autogenic effects on the site we studied would, if anything, be greater in the 20th century when evapotranspiring leaf area in the stand reached even greater levels than in the first few decades in the life of the stand. So while decreasing discrimination indicates increasing moisture stress, which may have been partly a result of less soil moisture being available to the trees as a result of their own growth, there is no reason to believe that the effect was uniquely strong in our trees in the mid 19th century. In addition, the amplitude of change in δ^{13}C discrimination in our sample appears to be greater than could be produced by the unique growth characteristics of a particular age class of trees (Bert et al., 1997). There is no systematic divergence in the relationship between δ^{13}C discrimination and Fairbanks climate data in the earliest years of the 20th century compared to the late 20th century, a nearly 100-year period during which either a decline or increase in leaf area would almost certainly have occurred. As a result, we believe that juvenile or autogenic effects on δ^{13}C discrimination are not the likely explanation for the reconstruction of summer warmth during Regimes 19.2A and 19.2C.

Hypothesis 2a, suggests that during the 19th century, before instrument-based climate records, summers were so cold that an ecological threshold was crossed and the relationship of Interior Alaska white spruce radial growth to summer temperature on sites similar to those we studied would have been positive. Under such a scenario, then, the nearly uniquely low radial growth during regimes 19.2A and 19.2C (Figure 5) would actually be a signal of cold summers. One obvious difficulty with this explanation is the magnitude of cooling necessary to produce an opposite temperature response. The 20th century mean July temperature at Fairbanks is 16.2 °C and the mean annual precipitation is 282 mm, placing this environment at the warm and dry margin of the occurrence of the species (Thompson et al., 2000). If the hypothesized cooling necessary to produce an opposite temperature sensitivity involved simply a move to the middle of the temperature tolerance range for the species with no change in precipitation, the magnitude of cooling would amount to about 3 °C in July temperature compared to the 20th century mean at Fairbanks (Thompson et al., 2000). July temperatures under such a scenario would thus be more than 2 °C cooler than Regime 20.1 (1916–1937), the coolest period in the instrument record. We offer additional data later in this discussion that appear to preclude such a cooling.

According to hypothesis 2b, it may have been cool to cold during the early to mid-19th century in Interior Alaska, but so dry that effective moisture was extremely low and therefore growth was limited. There is no modern analog in the nearly 100 years of recorded data for this type of climate (cool/dry). The two predominant climate modes in the 20th century recorded Fairbanks data are consistent with changes in overall circulation patterns that produce two prevalent anomalous summer climate patterns in Interior Alaska; strong maritime (cool and wet) versus continental (hot and dry) conditions. Achieving a pure precipitation signal from any of the few tree species in Alaska remains elusive, but it is clear that more work is needed to resolve the precipitation question. Of course, evidence that mid 19th century summer temperatures were not cold in Interior Alaska, but in fact were warm as we have reconstructed them, would eliminate this hypothesis as a plausible explanation. We present such evidence below.

4.3. ADDITIONAL TREE-RING DATA CONCERNING THE 19TH CENTURY CLIMATE RECONSTRUCTION

A sample of hybrid white/Sitka spruce (Lutz spruce) from Anchorage in south-central Alaska (FRA) maintains a positive radial growth response to Fairbanks summer temperature during the 20th century period of record (Figure 8). During the 19th century period of reconstructed temperature, radial growth in this sample shows two distinct periods of high radial growth during Regimes 19.2A and 19.2C. It is highly unlikely that this enhanced growth could have been achieved during a period of exceptional cold in the region, as would be required by hypothesis 2a. Although Anchorage is 420 km south of Fairbanks (Figure 2), warm season (May–August) temperature trends in both locations generally have been consistent during the period of instrument record (Juday, 1984), and during the common period (1916–1996) they correlate at 0.60.

A sample of white spruce at the North Fork of the Koyukuk River in the central Brooks Range (Figure 2) is a mixed population of trees with positive and negative radial growth responses to Fairbanks summer temperature (Figure 9). The

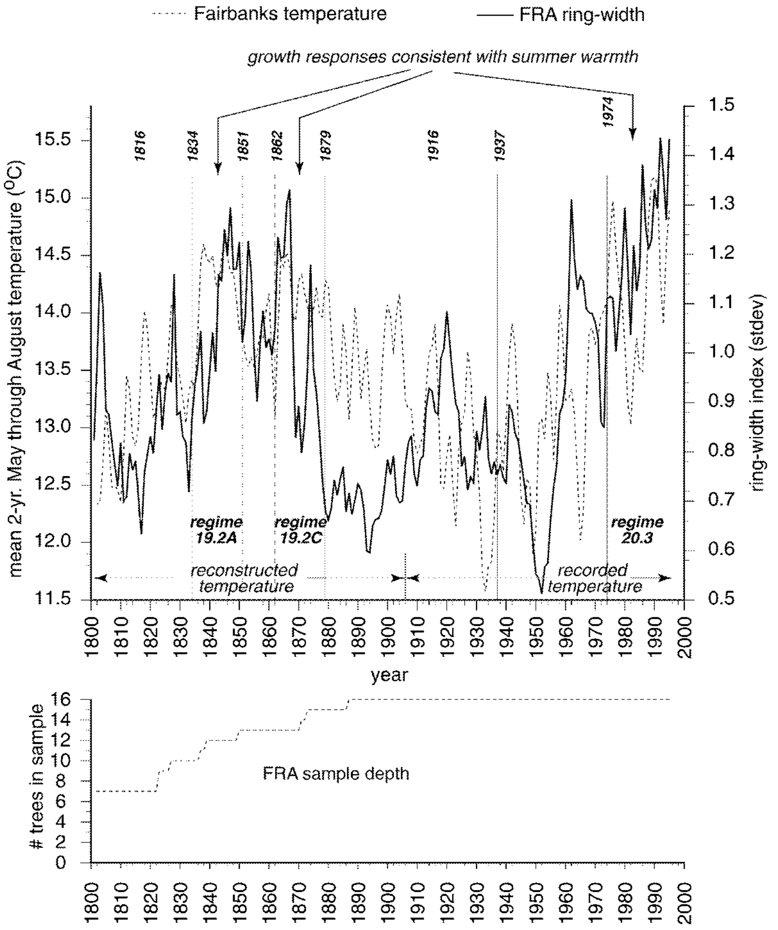

Figure 8. (top) Comparison of relative radial growth of white spruce since 1800 at Fort Richardson, Anchorage (FRA in Figure 2) versus composite recorded and reconstructed Fairbanks summer temperature (mean of growth year and prior year). Ring-width values represent smoothing (5-year running mean) of detrended, normalized transformation of ring-width, with mean set to 1.0. Correlation = 0.41, $p < 0.01$. Note strong positive growth responses during warm summers of Regime 20.3 (recorded) and Regimes 19.2A and 19.2C (reconstructed). Such positive growth responses in a majority of the sample (bottom) are difficult to account for except by optimum climatic conditions, which involve summer warmth in this population located in a humid coastal location.

trees with a negative growth response display a generally good match to Fairbanks temperatures throughout the entire 200 year period of this analysis, including two periods of low growth during Regimes 19.2A and 19.2C (Figure 9). Sample depth was not adequate to obtain a record continuously from 1800 onward for the positive responders, but the one tree with the longest record displays high radial growth during two reconstructed warm periods at Regimes 19.2A and 19.2C (Figure 9). The evidence of positive responding trees at contemporary tree limit displaying

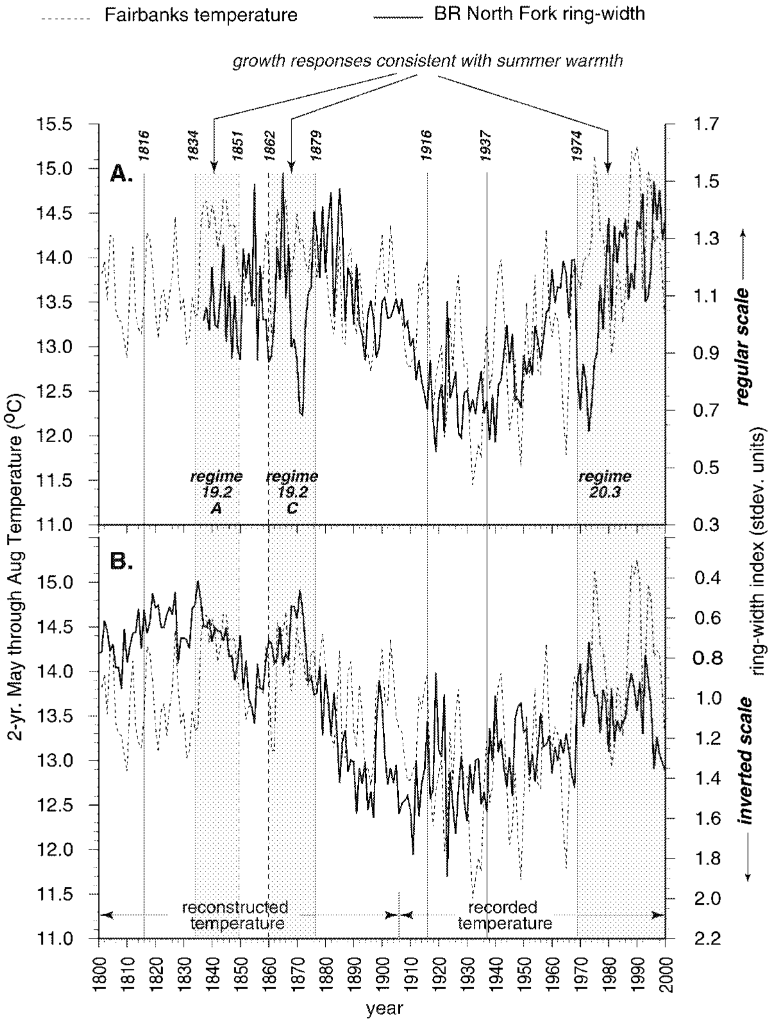

Figure 9. Comparison of relative radial growth of treeline white spruce since 1800 at North Fork in the central Brooks Range stand (location in Figure 2) versus composite recorded and reconstructed Fairbanks 2-yr. summer temperature (mean of growth year and prior year). Ring-width values are detrended, normalized transformations of ring-width, with mean = 1.0. A. Relationship of ring-width and 2-yr summer temperature in a population of trees with a positive response to temperature. Sample size starts at 2 in 1834 and increases to 4 with trees added at 1870 and 1873. Correlation = 0.36, $p < 0.01$. Note strong positive growth response during warm summers of Regime 20.3 (recorded). Relative tree growth is higher during much of the mid 19th century, including Regimes 19.2A and 19.2C, than nearly all of the late 19th century and all but the last decades of the 20th century. Such positive growth responses in this cold-limited population can be produced in practically no other way than warm summers. B. Relationship of ring-width and 2-yr. summer temperature in a population of trees with a negative response to temperature. Note inverted scale of ring-width. Sample size starts at 5 in 1800 and increases to 7 with trees added at 1817 and 1820. Correlation = –0.44, $p < 0.01$. Lowest period of growth during the 20th century occurred during Regime 20.3. Periods of low radial growth during Regimes 19.2A and 19.2C are consistent with warm temperatures.

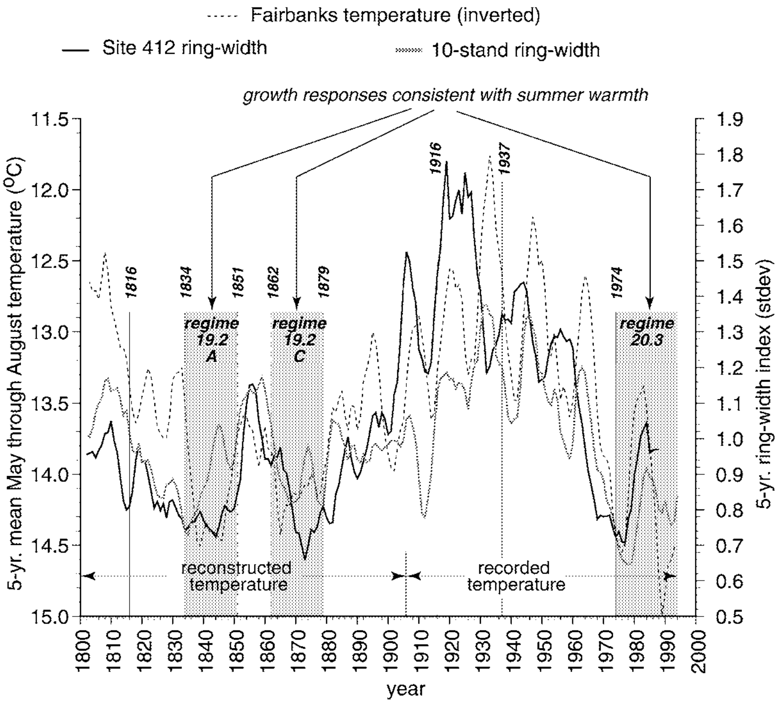

Figure 10. Comparison of smoothed (5-yr running mean) relative radial growth of white spruce since 1800 at 10-stand and Site 412 versus smoothed summer temperature at Fairbanks. Correlation of 10-stand to Fairbanks = –0.52, $p < 0.01$, Site 412 to Fairbanks = –0.59, $p < 0.01$. Because of negative correlation, temperature scale is inverted. Both radial growth indices are well correlated (negatively) with Fairbanks summer temperature through the period of comparison and show minimal radial growth during Regimes 20.3, 19.2A, and 19.2C, a response consistent with elevated summer temperatures.

high radial growth during the mid-19th century is particularly persuasive evidence indicating warm summer temperatures and does not support hypothesis 2a.

An important tree-ring data set used in the standard reconstruction of Arctic temperature variability (Overpeck et al., 1997) is the Site 412[⋆] (67°56′N, 162°18′W) sample of white spruce at tree limit in northwest Alaska. A comparison of (1) the composite reconstructed and recorded Fairbanks summer temperature, (2) mean growth of the 10-stand Interior Alaska white spruce sample, and (3) the Site 412 radial growth record display a high degree of agreement throughout the period from 1800 to the present (Figure 10). Notably, the 3 records are consistent during our reconstructed two mid-19th century warm periods in question.

Finally, another compelling piece of evidence comes from the growth record of a different species. At the Zasada Road 10 ridge top site in Bonanza Creek LTER near Fairbanks, the correlation of black spruce radial growth with Fairbanks monthly temperatures is highest for May in the year of ring formation and the

---

[⋆] (ftp://ftp.ngdc.noaa.gov/paleo/treering/chronologies/asciifiles/usawest/ak031.crn).

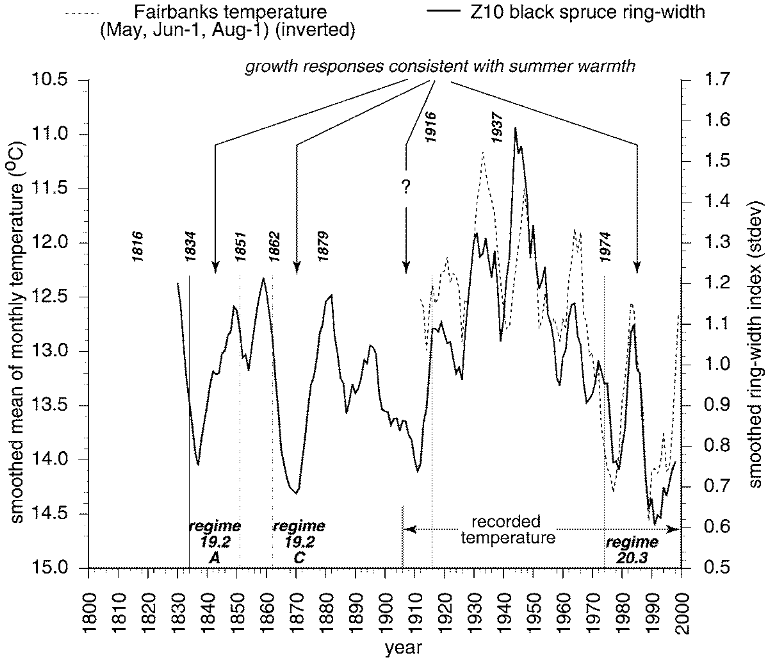

Figure 11. Comparison of smoothed (5-yr running mean) relative radial growth of black spruce since 1830 at Zasada Road 10 versus smoothed mean of recorded May, previous July, and previous August monthly temperatures at Fairbanks. Correlation = –0.81, $p < 0.01$. Because of negative correlation, temperature scale is inverted. Carbon-13 discrimination and maximum latewood density did not reconstruct this combination of monthly temperatures so we show no reconstruction for the 19th century, but smoothed (5-year mean) May through August temperature is highly correlated with smoothed May, previous July and previous August at 0.933, $p < 0.01$. Minimal periods of radial growth occur at Regimes 20.3, 19.2A, 19.2C, and the period 1898–1916, a response consistent with elevated summer temperatures.

previous July and August (Figure 11). The correlation is consistently negative throughout the period of instrument-based climate record. The mean of the three monthly temperatures that correlates with black spruce growth represents slightly different information content than the May through August reconstructed temperature mean, but the two temperature means correlate at 0.63 (annual values) during the period 1907–1996. Again, two periods of low black spruce radial growth are obvious during Regimes 19.2A and 19.2C (Figure 11) which we have reconstructed as warm periods.

These tree growth responses of (1) the same species (as our proxy sample) responding in a positive manner south of Interior Alaska, (2) the same species responding simultaneously in a positive and negative manner at northern tree limit, and (3) a different species responding in a negative manner at the same locality as the sample for the 2-proxy reconstruction are consistent throughout the period of recorded data and more importantly they are consistent with our reconstruction. In light of these supplemental records of tree growth, any explanations that require Regimes 19.2A and 19.2C to be something other than warm become implausibly complex. The simplest explanation is that the mid-19th century summers in Interior Alaska were indeed warm. While the alternative hypotheses represent possible explanations for our results, they are not supported by the additional data. Thus the weight of the evidence suggests that it is reasonable to conclude that our reconstructed summer temperature data represent some real differences between Interior Alaska and the rest of northern North America.

Some other important climate-related features of summer temperature and Interior Alaska tree growth are worth noting. The cool (and moist) climate during regimes 20.1 and 20.2 was so favorable for trees limited by temperature-driven drought (Barber et al., 2000) that these years represent the period of greatest relative radial growth of white spruce on similar sites in Interior Alaska over the past 200 years by a substantial margin. Reciprocal relationships in relative growth between positive and negative responding white and black spruce are well synchronized with temperature variability (Figures 8, 9, 10 and 11) (Juday, Barber, Rupp, Zasada and Wilmking, in press) suggesting that carbon uptake in the Alaska boreal forest almost certainly varies in accordance with our defined regimes according to the area-weighted extent of the respective species and site types.

5. Conclusions

Our multi-proxy 200-year summer temperature reconstruction is one of the first regional scale high-resolution reconstructions for Interior Alaska and shows excellent agreement with the recorded data from the 20th century. The reconstruction, as well as the recorded data, displays decadal-scale regimes, which appear to be the result of pervasive synoptic-scale climate patterns in the western North American boreal region.

Summer temperature trends in Interior Alaska over the past 200 years are different from the overall trend for the Northern Hemisphere, and from other Northern Hemisphere reconstructions for this time period. Specifically, we reconstruct summer temperatures during the two periods in the mid-1800s that are about as warm as present. While we cannot completely rule out autogenic factors playing some role in our reconstruction nor the possibility that the sensitivity of tree-ring properties may have been different in the 19th century, neither appears likely. Long-term radial growth responses of white spruce and black spruce do not support the hypothesis that the climate was cool to cold, and so dry during this time period, that moisture was limiting. The anomalies we have identified most likely reflect very real and different regional synoptic conditions between Interior Alaska and much of the rest of northern North America. Alaska, as a peninsular extension of North America to the west, is peculiarly susceptible to the influx of marine air moving north off the Bering Sea and North Pacific. When this pulse is strong during the warm season, summer in Interior Alaska is anomalously cool and moist. When a blocking high pressure centered on Interior Alaska dominates circulation, the long days near the summer solstice cause hot and dry conditions. Alternation between these two modes of atmospheric circulation dominated the 20th century. Cool/wet conditions dominated from the early century until a shift to warm/dry conditions in the 1970s.

The warm summers during Regime 20.3 are unprecedented over the nearly 100-year period of recorded data, but not so in the perspective of the last 200 years. This is not to say that our data argue against human-induced global warming, but they appear to demonstrate the strong control of climate in Interior Alaska by changes in atmospheric circulation. Although it is probable that synoptic conditions similar to those in recent decades may have existed in the early to mid-19th century, additional evidence in the form of other proxies and examination of temperatures in different high latitude regions at synchronous periods may be able to further clarify the occurrence of warm season temperature anomalies that we have identified in central Alaska.

Acknowledgements

Financial support came from the McIntire-Stennis Cooperative Forestry Research Program, the University of Alaska's Natural Resource Fund and Center for Global Change, the US NSF Long-term Ecological Research (LTER) Program, and the U.S. DOE WESTGEC Program. We thank 3 anonymous reviewers and Terry Chapin for useful and important suggestions.

References

Anderson, P. M., Bartlein, P. J., Brubaker, L. B., Gajewski, K., and Ritchie, J. C.: 1991, 'Vegetation-Pollen-Climate Relationships for the Arcto-Boreal Region of North America and Greenland', *J. Biogeogr.* **18**, 565–582.

Barber, V. A., Juday, G. P., and Finney, B. P.: 2000, 'Reduced Growth of Alaskan White Spruce in the Twentieth Century from Temperature-Induced Drought Stress', *Nature* **405**, 668–673.

Barry, R. G. and Hare, F. K.: 1974, 'Arctic Climate', in Ives J. D. and Barry, R. G. (eds.), *Arctic and Alpine Environments*, 999 pp.

Bert, D., Leavitt, S. W., and Dupouey, J.-L.: 1997, 'Variations of Wood δ^{13}C and Water-Use Efficiency of Abies Alba during the Last Century', *Ecology* **78**, 1588–1596.

Bradley, R. S. and Jones, P. D.: 1993, 'Little Ice Age' Summer Temperature Variations: Their Nature and Relevance to Recent Global Warming Trends', *Holocene* **3**, 367–376.

Briffa, K. R., Schweingruber, F. H., Jones, P. D., Osborn, T. J., Shiyatov, S. G., and Vagnov, E. A.: 1998, 'Reduced Sensitivity of Recent Tree-Growth to Temperature at High Northern Latitudes', *Nature* **391**, 678–682.

Brooks, J. R., Flanagan, L. B., and Ehleringer, J. R.: 1998, 'Responses of Boreal Conifers to Climate Fluctuations: Indications from Tree-Ring Widths and Carbon Isotope Analyses', *Can. J. Forest Res.* **28**, 524–533.

Chapman, W. L. and Walsh, J. E.: 1993, 'Recent Variations of Sea Ice and Air Temperature in High Latitudes', *Bull. Amer. Meteorol. Soc.* **74**, 33–47.

Cook, E. R. and Kairiukstis, L. A.: 1990, *Methods of Dendrochronlogy: Applications in the Environmental Sciences*, Kluwer Academic Publishers, 394 pp.

D'Arrigo, R. D. and Jacoby, G. C.: 1993, 'Tree Growth Climate Relationships at the Northern Boreal Forest Tree Line of North America: Evaluation of Potential Response to Increasing Carbon Dioxide', *Global Biogeochem. Cycles* **7**, 525–535.

Edwards, M. E., Mock, C. L., Finney, B. P., Barber, V. A., and Bartlein, P. J.: 2001, 'Potential Analogues for Paleoclimatic Variations in Astern Interior Alaska during the Past 14,000 Yr: Atmospheric-Circulation Controls of Regional Temperature and Moisture Responses', *Quatern. Sci. Rev.* **20**, 189–202.

Francey, R. J., Allison, C. E., Etheridge, D. M., Trudinger, C. M., Enting, I. G., Leuenberger, M., Langenfelds, R. L., Michel, E., and Steele, L. P.: 1999, 'A 1000-Year High Precision Record of δ13C in Atmospheric CO_2', *Tellus* **51B**, 170–193.

Francey, R. J. and Farquhar, G. D.: 1982, 'An Explanation of 13C/12C Variations in Tree Rings', *Nature* **297**, 28–31.

Fritts, H. C.: 1976, *Tree-Rings and Climate*, Academic Press, 567 pp.

Garfinkle, H. L. and Brubaker, L. B.: 1980, 'Modern Climate-Tree-Growth Relationships and Climatic Reconstruction in Sub-Arctic Alaska', *Nature*, **286**, 872–874.

Hogg, E. H.: 1994, 'Climate and the Southern Limit of the Western Canadian Boreal Forest', *Can. J. Forest Res.* **24**, 1835–1845.

Hogg, E. H.: 1997, 'Temporal Scaling of Moisture and the Forest-Grassland Boundary in Western Canada', *Agric. For. Meteorol.* **84**, 115–122.

Holmes, R. L.: 2000, Cofecha. Laboratory of Tree-Ring Research.

Houghton, J. J., Meiro Filho, L. G., Callander, B. A., Harris, N., Kattenberg, A., and Maskell, K. (eds.): 1996, *The Science of Climate Change. Contribution of Working Group I to the Second Assessment Report of the Intergovernmental Panel on Climate Change (IPCC), Vol. 1, Climate Change 1995*, Cambridge University Press, 584 pp.

Hughes, M. K. and Diaz, H. F.: 1994, 'Was There a "Medieval Warm Period", and if so, Where and When?', *Clim. Change* **26**, 109–142.

Jacoby, G., Lovelius, N., Shumilov, O., Raspopov, O., Kabainov, J., and Frank, D.: 2000, 'Long-Tern Temperature Trends and Tree Growth in the Taymir Region of Northern Siberia', *Quatern. Res.* **53**, 312–318.

Jacoby, G. C. and D'Arrigo, R.: 1989, 'Reconstructed Northern Hemisphere Annual Temperature since 1671 Based on High-Latitude Tree-Ring Data from North America', *Clim. Change* **14**, 39–59.

Jacoby, G. C. and D'Arrigo, R. D.: 1995, 'Tree Ring Width and Density Evidence of Climatic and Potential Forest Change in Alaska', *Global Biogeochem. Cycles* **9** 227–234.

Jacoby, G. C., D'Arrigo, R. D., and Davaajamts, T.: 1996, 'Mongolian Tree Rings and 20th-Century Warming', *Science* **273**, 771–773.

Johnson, C. A., Pastor, J., and Pinay, G.: 1992, 'Quantitative Methods for Studying Landscape Boundaries', in Hansen, A. J. and di Castri, F. (eds.), *Landscape Boundaries: Consequences for Biotic Diversity and Ecological Flows*, Springer-Verlag, 452 pp.

Juday, G. P.: 1984, 'Temperature Trends in the Alaska Climate Record', in *Proceedings of the Conference on the Potential Effects of Carbon Dioxide-Induced Climatic Changes in Alaska*, Fairbanks, AK, Ag. Exp. Sta. Misc. Publication, Univ. of Alaska, pp. 76–88.

Kasischke, E. S. and Stocks, B. J.: 2000, *Fire, Climate Change, and Carbon Cycling in the Boreal Forest*, Springer-Verlag, 461 pp.

Koskey, M. S. and Yamin, S., 2001: 'Climate Variability in the Bering Strait Region: Written Sources and the Detection of Arctic Climate Change', *Global Glimpses*, 5–6.

Kurz, W. A., Apps, M. J., Stocks, B. J., and Volney, W. J. A.: 1995, 'Global Climate Change; Disturbance Regimes and Biospheric Feedbacks of Temperature and Boreal Rorests', in *Boreal Forests and Global Change*, Kluwer Academic Publishers, 548 pp.

Labau, V. J. and Van Hees, W. W. S.: 1990, 'An Inventory of Alaska's Boreal Forests: Their Extent, Condition, and Potential Use', in *Proceedings International Symposium Boreal Forests: Climate, Dynamics, Anthropogenic Effects*, Archangelsk, Russia, State Committee of U.S.S.R. on Forests.

Leavitt, S. W. and Danzer, S. R.: 1992, 'Method for Batch Processing Small Wood Samples to Holocellulose for Stable-Carbon Isotope Analysis', *Analytical Chemistry* **65**, 87–89.

Lloyd, A. H. and Fastie, C. L.: 2002, 'Spatial and Temporal Variability in the Growth and Climate Response of Treeline Trees in Alaska', *Clim. Change* **52**, 481–509.

Mann, M. E., Bradley, R. S., and Hughes, M. K.: 1998, 'Global-Scale Temperature Patterns and Climate Forcing over the Past Six Centuries', *Nature* **392**, 698–702.

Mock, C. J., Bartlein, P. J., and Anderson, P. M.: 1998, 'Atmospheric Circulation Patterns and Spatial Climatic Variations in Beringia', *Int. J. Clim.* **18**, 1085–1104.

Nienstaedt and Zasada, J.: 1990, '*Picea Glauca* (Moench) Voss White Spruce', in Burns, R. M. and Honkala, B. H. (eds.), *Silvics of North America: 1. Conifers; 2. Hardwoods*, U.S. Department of Agriculture, Forest Service, 877 pp.

Osterkamp, T. E.: 1996, 'Impacts of Thawing Permafrost as a Result of Climatic Warming', *EOS Trans. Am. Geophys. Union* **77**, F188.

Osterkamp, T. E. and Romanovsky, V. E.: 1999, 'Evidence for Warming and Thawing of Discontinuous Permafrost in Alaska', *Permafrost and Periglacial Processes* **10**, 17–37.

Overpeck, J., Hughen, K., Hardy, D., Bradley, R., Case, R., Douglas, M., Finney, B., Gajewski, K., Jacoby, G., Jennings, A., Lamoureux, S., Lasca, A., MacDonald, G., Moore, J., Retelle, M., Smith, S., Wolfe, A., and Zielinski, G.: 1997, 'Arctic Environmental Change of the Last Four Centuries', *Science* **278,** 1251–1256.

Patric, J. H. and Black, P. E.: 1968, 'Potential Evapotranspiration and Climate in Alaska by Thornthwaite's Classification', Research Paper PNW-71, U.S.D.A. Forest Service Juneau, AK.

Ruess, R. W., Van Cleve, K., Yarie, J., and Viereck, L. A.: 1996, 'Comparative Estimates of Fine Root Production in Successional Taiga Forests on the Alaskan Interior', *Can. J. Forest Res.* **26**, 1326–1336.

Salmon, D. K.: 1992, *On Interannual Variability and Climate Change in the North Pacific (Wind Stress Curl)*, Ph.D. Institute of Marine Science, University of Alaska Fairbanks, 229 pp.

Serreze, M. C., Walsh, J., Chapin, F. S., Osterkamp, T., Dyurgerov, M., Romanosky, V., Oechel, W., Morison, J., Zhang, T., and Barry, R.: 2000, 'Observational Evidence of Recent Change in the Northern High-Latitude Environment', *Clim. Change* **46**, 159–207.

Thompson, R. S., Anderson, K. H., and Bartlein, P. J.: 2000, 'Atlas of Relation between Climatic Parameters and Distributions of Important Trees and Shrubs in North America – Introduction and Conifers', U.S. Geological Survey Professional Paper 1650-A, U.S. Geological Survey.

Wadhams, P.: 1995, 'Arctic Sea Ice Extent and Thickness', *Phil. Trans. Roy. Soc. London A.* **352**, 301–318.

Werner, R. A.: 1996, 'Forest Health in Boreal Ecosystems of Alaska', *The Forestry Chronicle* **72**, 43–46.

Wiles, G. C. and Calkin, P. E.: 1990, 'Neoglaciation in the Southern Kenai Mountains, Alaska', *J. Glaciol.* **14**, 319–322.

Wiles, G. C., D'Arrigo, R. D., and Jacoby, G. C.: 1996, 'Temperature Changes along the Gulf of Alaska and the Pacific Northwest Coast Modeled from Coastal Tree Rings', *Can. J. Forest Res.* textbf26, 474–481.

(Received 1 February 2002; in revised form 31 March 2003)