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VEGETATION MEDIATED THE IMPACTS OF POSTGLACIAL CLIMATIC CHANGE ON FIRE REGIMES IN THE SOUTHCENTRAL BROOKS RANGE, ALASKA

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ABSTRACT

We examine direct and indirect impacts of millennial-scale climatic change on fire regimes in the southcentral Brooks Range, Alaska, using four lake-sediment records and existing paleoclimate interpretations. New techniques are introduced to identify charcoal peaks semi-objectively and detect statistical differences in fire regimes. Peaks in charcoal accumulation rates (CHARs) provide estimates of fire return intervals (FRIs) which are compared between vegetation zones described by fossil pollen and stomata. Climatic warming from ca 15-9 ka BP (calendar years before CE 1950) coincides with shifts in vegetation from herb tundra to shrub tundra to deciduous woodlands, all novel species assemblages relative to modern vegetation. Two sites cover this period and show increased CHARs and decreased FRIs with the transition from herb to shrub tundra ca 13.3-14.3 ka BP. Short FRIs in the *Betula*-dominated shrub tundra (mean [m] FRI 144 yr; 95% CI 119-170) primarily reflect the effects of flammable, continuous fuels on the fire regime. FRIs increased significantly with the transition to *Populus*-dominated deciduous woodlands ca 10.5 ka BP (mFRI 251 yr [158-352]), despite evidence of warmer- and drier-than-present summers. We attribute reduced fire activity under these conditions to low flammability of deciduous fuels. Three sites record the mid to late Holocene, when cooler and moister conditions allowed *Picea glauca* forest-tundra and *P. mariana* boreal forests to establish ca 8 and 5.5 ka BP. Forest-tundra FRIs did not differ significantly from the previous period (mFRIs range from 131-238 yr), but FRIs decreased with the transition to boreal forest (mFRI 145 yr [129-163]). Overall, fire-regime shifts in the study area showed greater correspondence with vegetation characteristics than with inferred climate, and we conclude that vegetation mediated the impacts of millennial-scale climatic change on fire regimes by modifying landscape

flammability. Our findings emphasize the importance of biological-physical feedbacks in determining the response of arctic and subarctic ecosystems to past, and by inference, future climatic change.

Keywords: Alaska; arctic; boreal forest; charcoal analysis; climatic change; deciduous woodland; fire history; pollen analysis; shrub tundra; tundra.

INTRODUCTION

Recent warming in northern high latitudes (Overpeck et al. 1997, Serreze et al. 2000) has initiated a variety of changes in vegetation and fire regimes, including population expansion and increased growth of trees and shrubs (Lloyd 2005, Tape et al. 2006) and increased area burned across boreal forests (Kasischke and Turetsky 2006, Soja et al. 2007). As climate warms over the next century (ACIA 2004), annual area burned in boreal forests is predicted to continue increasing (Rupp et al. 2000, Calef et al. 2005), in some cases by more than 100% (Flannigan et al. 2005). The response of fire regimes to climatic change in these regions will ultimately depend upon feedbacks between climate and vegetation that determine landscape flammability at multiple temporal scales.

Numerous studies in North American boreal forests document the importance of direct climatic controls on fire regimes. At annual time scales, area burned in recent decades has been tightly linked to warm, dry weather and frequent lightning (Johnson 1992, Kasischke et al. 2002, Duffy et al. 2005). Paleoecological studies support these climate-fire relationships over millennial time scales and illustrate that fire frequencies have varied with changes in relative moisture throughout the late Holocene (Carcaillet et al. 2001, Lynch et al. 2002, Lynch et al.

2004b). However, climatic influences on fire regimes are also mediated by vegetation. Area burned across interior Alaska, for example, is positively correlated with tree cover (Kasischke et al. 2002), and the probability of fire in many boreal forests varies with stand age, a determinant of fuel abundance (Yarie 1981, Lynch et al. 2002). In addition, modeling studies predict increased burning in boreal Alaska simply in response to increased black spruce (*Picea mariana* Mill. BSP.) densities (Rupp et al. 2002).

We use a paleoecological approach to examine the relationships between climate, vegetation, and fire regimes in the southcentral Brooks Range, Alaska (Fig. 1), where millennial-scale climate and vegetation histories are well known based several decades of research in the region (Anderson et al. 2004). Our goal is to document fire history to understand how millennial-scale climatic changes have interacted with vegetation change to influence fire regimes over the past 15,000 years. We use macroscopic charcoal from lake sediments to reconstruct fire occurrence and statistically compare fire return intervals (FRIs, the time between consecutive fires) between vegetation zones inferred from fossil pollen, stomata stratigraphy, and modern analog analysis. If climatic variations were the dominant control of fire regimes, changes in fire occurrence between vegetation zones should be consistent with direct climate-fire relationships and relatively independent of vegetation characteristics (e.g. Carcaillet et al. 2001). In contrast, if vegetation change was the dominant control of fire regimes, changes in fire occurrence between vegetation zones should be consistent with the role that fuels play in determining landscape flammability (e.g. Lynch et al. 2002), but possibly unexpected given the direct effects of climate change on fire regimes. In addition to testing these two alternative hypothesis, by documenting fire regimes in novel vegetation assemblages that covered the study region during late glaciation and the early Holocene (c. 14.0-9.0 k ybp; e.g. Anderson et al. 1989, Edwards et al. 2005), our study provides examples of how direct and indirect impacts of climatic change may shape future fire regimes in arctic and boreal ecosystems.

Study Lakes and Regional Setting

We examined sediment cores from four lakes along a 120 km east-west transect in the foothills of the southcentral Brooks Range (Table 1, Fig. 1). Modern climate in the study region is continental. January and July mean maximum temperatures in Bettles (Fig. 1) are -20°C and 21°C, respectively; mean annual precipitation is 36 cm, with 55% falling between June and September (Western Regional Climate Center, 1951-2005 observations[1]). Forests and woodlands dominate lowlands and hill slopes in the study region, with *Picea mariana* in wet muskegs, *P. glauca* (Moench) Voss. and *Populus balsamifera* Mill. along riparian areas, and *P. glauca*, *Betula papyrifera* Marsh. and *Populus tremuloides* Michx. on uplands and warm, south-facing slopes (Nowacki et al. 2000). *Salix* spp., *Betula glandulosa* Michx., and *Alnus* spp. form shrub communities in non-forested areas, (Nowacki et al. 2000). Fire is the primary disturbance agent in the region, with an estimated fire rotation period of 175 yr (based on observations from 1950-2001 from the Kobuk Ridges and Valleys Ecoregion; Kasischke et al. 2002).

We cored small (2-15 ha), relatively deep (7.0-11.6 m) lakes (Table 1) surrounded by discontinuous *P. mariana*-dominated forest. Recent fires burned to the edge of Ruppert Lake (RP) in CE 1991 (15,357 ha), to 1 and 3 km east of Code Lake (CO) in CE 1959 (788 ha) and 1949 (2456 ha), and to 5 km west and 1 km southwest of Wild Tussock Lake (WK) in CE 1997 (9750 ha) and 1991 (6390 ha; Alaska Fire Service 2004; Fig. 1).

[1] Data available online: http://www.wrcc.dri.edu/cgi-bin/cliMAIN.pl?akbett

METHODS

Lake sediments

Sediments were collected from the center of each lake with two parallel, overlapping 8-cm diameter cores in summer 2001 (CO), 2002 (RP), or 2003 (XI, WK) using a modified Livingstone piston corer (Wright et al. 1984). Surface sediments (< ca 50 cm) were collected with a polycarbonate tube and the top 10-20 cm sliced at 0.5-1.0 cm in the field. All cores had intermittent laminae, and stratigraphic markers were used to develop continuous records from overlapping segments of adjacent cores. Cores were sliced at 0.25 to 0.50 cm intervals, and 1 cm^3 subsamples were prepared at varying intervals for pollen and stomata analysis according to PALE (1994), except that samples were not subjected to a coarse sieve (Carlson 2003). Pollen was counted at 400-1000 x magnification to a terrestrial pollen sum > 300 (mean = 398, s.d. = 107) and displayed as percentages of total terrestrial pollen. In samples bracketing the *Picea* pollen rise ca 5.5 ka BP (Brubaker et al. 1983), (1) stomata searches were conducted to an equivalent pollen sum of 2000 grains (Carlson 2003) and *Picea* stomata were identified based on comparisons with an Alaskan reference collection and Hansen (1994), and (2) *Picea* pollen grains were classified as *P. mariana* or *P. glauca* based on morphological measurements on ca 30 *Picea* pollen grains per sample (Appendix A). For charcoal identification, 3-5 cm^3 subsamples were taken from contiguous core slices and prepared following Higuera et al. (2005). Charcoal was identified at 10-40 x magnification based on color, morphology, and texture. Charcoal concentrations (pieces cm^{-3}) were multiplied by the estimated sedimentation rate (cm yr^{-1}) to obtain the charcoal accumulation rate (CHAR; pieces cm^{-2} yr^{-1}) of each sample.

Chronologies

Sediment chronologies for each site were based on ^{210}Pb dates for the upper 10-20 cm (using the CRS model; Binford 1990) and/or on AMS ^{14}C ages of concentrated charcoal from charcoal peaks, concentrated *Picea* pollen, or terrestrial macrofossils for deeper sediments. Chronologies were developed individually for the ^{210}Pb and ^{14}C portions of each core using a weighted cubic smoothing spline in Matlab (The MathWorks, Inc.) taking into account the number and uncertainty of age estimates. The number of age estimates in a chronology determined the smoothing parameter for each spline, such that a larger number of ages for a given time interval resulted in a more flexible spline. The uncertainty of each age estimate (i.e. two standard deviations) was used to weight the influence of each age in the age-depth model (cf. Telford et al. 2004). Finally, confidence intervals for the age-depth model, reflecting the combined uncertainty of all age estimates in a model, were derived from 1000 bootstrapped chronologies. For each bootstrapped chronology, each age used to develop the chronology was selected randomly based on the probably distribution of the ^{210}Pb or calibrated ^{14}C date. The final chronology represents the median age at each depth from the 1000 bootstrapped chronologies.

Analysis of pollen and stomata data

Pollen zone boundaries were delineated primarily by visual inspection of pollen percentages of major tree, shrub, and herb taxa and correspond to vegetation types previously recognized in the region (Anderson and Brubaker 1994). In addition, we used a modern analog analysis based on squared-chord distances (SCD) and receiver operating characteristic curves to quantify the probability that fossil pollen assemblages resembled modern pollen assemblages from North American Arctic Tundra, Boreal Forest, and Forest-tundra biomes (Appendix A). The arrival of *P. mariana* and development of the modern boreal forest was estimated via (1) *Picea* stomata presence/absence (RP and WK), (2) discriminant analysis (RP and WK) of *Picea* grains, and (3) modern analog analysis (Appendix A).

Statistical treatment of charcoal records

To assess whether CHARs differed between past biomes, we compared CHAR distributions between pollen zones in each lake using a two-sample Kolmogorov-Smirnov (K-S) test (e.g. Clark 1990, Lynch et al. 2002). Prior to subsequent analyses, which emphasize finer temporal variation in CHARs, we interpolated CHARs to 15-yr time steps ($C_{interpolated}$), approximating the mean sampling resolution at all sites (Table 1). This step reduces biases due to variable sample resolution within and between records.

We inferred the timing of “local” fire occurrence by decomposing charcoal records (e.g. Clark et al. 1996) and identifying “large”, distinct charcoal peaks based on a standard set of threshold criteria applied to each record. Consistent with empirical and theoretical studies (Lynch et al. 2004a, Higuera et al. 2007), we use the term “local” to refer to distances within approximately 500-1000 m of each lake, corresponding to an area of ca 100-300 ha (1-3 km^2). We assume that low-frequency variations in CHARs, $C_{background}$, reflect changes in the rate of total charcoal production, secondary charcoal transport, and sediment mixing (Clark et al. 1996, Higuera et al. 2007). We subtracted $C_{background}$ from $C_{interpolated}$ to obtain a residual series, C_{peak} (i.e. $C_{peak} = C_{interpolated} - C_{background}$), which contains high-frequency variability around the long-term trend described by $C_{background}$ (e.g. Clark et al. 1996).[2] Prior decomposition methods have assumed a constant relationship between C_{peak} and $C_{background}$ and applied a globally-defined threshold values to C_{peak} series to identify peaks that reflect the occurrence of one or more local fires. In our records, however, the variability of C_{peak} around $C_{background}$ changes through time (Appendix A). Thus, we developed a locally-defined threshold that identifies charcoal peaks based on local variability around each sample. Specifically, we used a Gaussian mixture model to separate C_{peak} values within each overlapping 500-yr portion of a record into two components: (1) C_{noise}, variations around $C_{background}$ that reflect natural and analytical effects (e.g. sediment mixing, sediment sampling), and (2) C_{fire}, variations exceeding variability in the C_{noise} distribution, assumed to reflect local fire events (Gavin et al. 2006). The threshold separating C_{fire} from C_{noise} should occur in the upper range of the C_{noise} population. Thus, to identify C_{fire} samples in any 500-yr portion of the record we considered three possible threshold values corresponding to the 95th, 99th, and 99.9th percentile of the C_{noise} population. We present results from all three criteria but discuss only those using the 99th percentile criterion. Finally, all peaks identified via these methods were screened to test whether variations between a “peak” and the smallest “non-peak” sample within the previous 5 samples (i.e. 75 yr) differed statistical based on the size of the original charcoal counts (Gavin et al. 2006; Appendix A). This screening eliminates “peaks” during periods of low charcoal accumulation (e.g. < 5 pieces per sample) when the variation in low charcoal counts cannot be confidently separated from noise. Our

[2] We estimated low-frequency trends by calculating a 500-yr moving median across each record and then smoothing this series with a locally-weighted regression using a 500-yr window.

methods are described in detail in Appendix A and are contained within the publicly-available program *CharAnalysis*[3].

In addition, as an indication of the suitability of charcoal records for peak identification, we calculated a signal-to-noise index (*SNI*) for each sample, SNI_i, describing the variance within the C_{fire} distribution (i.e. signal) relative to the total variance in C_{peak} in the 500 years surrounding that sample: $SNI_i = \frac{\mathrm{var}(C_{fire})}{[\mathrm{var}(C_{noise}) + \mathrm{var}(C_{fire})]}$ The SNI varies from 0-1, with high values representing large separation between charcoal peaks and non-peaks, and values close to 0 representing little separation between peaks and non peaks.

Quantifying and detecting difference in fire regimes

We infer aspects of past fire regimes based on the magnitude and temporal pattern of identified charcoal peaks. Peak magnitude, the number of charcoal pieces from all samples defining a peak (i.e. all samples above the threshold value; pieces cm^{-2} peak^{-1}), is measure of total charcoal deposition per fire (Whitlock et al. 2006). Charcoal deposition for a given fire should vary depending on fire size, fire location, and fuel consumption (Higuera et al. 2007). Systematic changes in peak magnitude thus reflect variability in fuel consumption, as variability due to fire size and fire location is accounted for as multiple fires are sampled.

We used the distribution of FRIs within each pollen zone to characterize the FRIs of each vegetation type. FRI distributions were described by the mean FRI (mFRI), and if a pollen zone had > 5 FRIs (> 6 fires), a two-parameter Weibull model was fit to FRIs using maximum likelihood techniques (in Matlab, The Mathworks Inc.; Clark 1989, Johnson and Gutsell 1994).

[3] *CharAnalysis* was written by PEH and is available at http://CharAnalysis.googlepages.com

Goodness-of-fit for each Weibull model was tested with a one-sample K-S test (Zar 1999) and Weibull models are not reported unless $p > 0.10$ (i.e. there is > 10% chance that the empirical distribution is not different from the Weibull model; Johnson and Gutsell 1994). 95% confidence intervals for Weibull parameters and mFRIs were estimated based on 1000 bootstrapped samples from each distribution. We tested two null hypotheses using a likelihood-ratio test (LRT) [4] based on estimates of the Weibull *b* and *c* parameters (Appendix A): (1) FRIs did not differ between pollen zones within a given site, and (2) FRIs did not differ between sites within a given pollen zone. We rejected the null hypothesis if $p \leq 0.05$. If FRI distributions within a single vegetation zone were statistically similar across sites (i.e. hypothesis 2 was not rejected), we pooled FRIs to form a composite record representing FRI from across the study area (i.e. 2-3 sites). With the pooled FRIs we performed the same between-zone comparisons as with individual records. Due to the increased sample size in the composite records, this procedure yields greater statistical power for detecting differences between fire regimes in different vegetation zones. Because in most cases comparisons between zones at individual sites were similar to results for the pooled data (although not always significant), we focus our discussion on results from the pooled data.

RESULTS

Chronologies and sedimentation rates

The RP and XI records are older (14 and 15.5 ka) than the CO and WK records (7.5 and 7.8 ka; Fig. 2). All age models since 8 ka BP are well constrained and generally pass through the

[4] By utilizing both parameters of the Weibull distribution, the LRT provides a more powerful method for detecting difference in FRI distributions than possible by interpreting confidence intervals around mFRI and estimated Weibull parameters (e.g. Clark 1990, Lynch et al. 2002) or by using the non-parametric K-S test (Lynch et al. 2002, Anderson et al. 2006, Gavin et al. 2006).

95% confidence interval of ^{14}C or ^{210}Pb dates (Fig. 2). At RP we rejected two ^{14}C dates on concentrated pollen (19.02 and 29.50 cm) because they are ca 500-1000 yr older than ages defined by five other ^{14}C dates on charcoal in sediments spanning the same core depths (10 and 60 cm; Fig. 2; Appendix B). At RP and XI, age models > 8 ka BP are less well constrained, and predicted ages do not always intersect the uncertainty of ^{14}C dates (e.g. RP; Fig. 2). Given these results and the sensitivity of CHARs to sedimentation rates, we evaluated whether different choices of age-depth relationships altered the general features of the CHAR series at these sites. We developed 5-7 alternative age-depth relationships by excluding individual dates and changing the age-depth model criteria. In no case did the overall nature of the CHAR records change.

Sedimentation rates ranged from 0.017 to 0.040 cm yr^{-1} (Table 1, Fig. 2). They vary little in the CO and WK records, but are higher prior to ca 8 and 11 ka BP at RP and XI, respectively (Fig. 2). The mean sample resolution at each site is ~15 yr $sample^{-1}$ (Table 1), varying from < 5 to ca 50 yr $sample^{-1}$ (Fig. 2). Slow sedimentation resulted in low sample resolution at XI from 8.0-0 ka BP (> 50 yr $sample^{-1}$).

Peak identification in charcoal records

. The median signal-to-noise index (SNI) for all records where peak analysis was interpreted exceeds 0.80 (Table 1). This is well above the median SNI expected from records without a peak signal (e.g. 0.15 for red noise; Appendix A) and indicates good separation between peak and non-peak values. We do not interpret peak analysis results at XI from 8.0-0 ka BP because the SNI in this period is consistently << 0.5 (data not shown). The sensitivity of charcoal peak identification to different threshold criteria varies between pollen zones (Fig. 3;

Appendix B) and between sites (Fig. 4) but characterizations of FRI distributions are generally robust to all three threshold criteria (data not shown).

Comparisons between known fire events and the most recent charcoal peaks at RP, CO, and WK support the assumption that identified charcoal peaks detect fires within (and not beyond) 1 km of these lakes. The CE 1991 (-41 yr BP) fire that burned to the edge of RP is represented by an identified peak centered at -39 BP, while the most recent peaks identified at CO (69 yr BP) and WK (38 yr BP) both occur before the start of fire observation in CE 1950 (0 yr BP); thus the fires that burned to ca 1, 3, and 5 km from these lakes were undetected.

Pollen, stomata, and charcoal records

Herb Tundra Zone: 14.0-13.3 (RP), 15.5-14.3 (XI) ka BP – This zone is characterized by Cyperaceae (>25%), *Salix* (ca 25%), Poaceae (ca 15%), and *Artemisia* (ca 10%) pollen, with relatively high percentages of *Pediastrum* algal cell nets (> 25%; Fig. 3; Appendix B). Although SCDs are lowest for comparisons with Arctic Tundra (ca 0.2), the probability-of-analog (< 20%) indicates little similarity with modern tundra (Fig 3; Appendix B). CHARs are low at both sites (medians = 0.00-0.01 pieces cm-2 yr-1; Appendix B). The presence of only one identified charcoal peak (at RP; Fig. 4) precludes the analysis of fire regimes but suggests long FRIs in this zone.

Shrub Tundra Zone, 13.3-10.3 (RP), 14.3-10.3 (XI) ka BP -- Increased *Betula* pollen (*B. glandulosa* and/or *B. nana*; Anderson and Brubaker, 1994) to > 60% marks the transition from herb to shrub tundra (Fig. 3; Appendix B). SCDs (ca 0.2) continue to indicate a low similarity of fossil to modern pollen assemblages from all modern biomes (probability of analog < 20%; Fig. 3; Appendix B). CHARs increase at the onset of this zone (medians = 0.02-0.05 pieces cm-2 yr-

1), and CHAR distributions are distinct from those in the Herb Tundra Zone ($p << 0.01$; Appendix B). Maximum peak magnitudes exceed 5 pieces cm^{-2} $peak^{-1}$ (Fig. 4). Fire regimes at RP and XI are characterized by mFRIs (95% CI) of 137 (107-171) and 150 (113-189) yr, respectively (Table 2; Fig. 5), with no difference in FRI distributions between sites (Fig. 5; Appendix B). The mFRI (95% CI) of the pooled record is 144 (119-170) yr (Fig. 6).

Deciduous Woodland Zone, 10.3-8.5 (RP), 10.3-8.0 (XI) ka BP – This zone is characterized by increased *Populus* pollen percentages (inferred as *P. balsamifera*, Anderson and Brubaker, 1994; 10-20%; Fig. 3; Appendix B). SCDs are the highest for the entire record (> 0.3), and no analogs exist with modern North America pollen spectra (probability of analog < 0.2; Fig. 3; Appendix B). CHARs decrease (medians = 0.01-0.02 pieces cm^{-2} yr^{-1}) and distributions are distinct from those in the Shrub Tundra Zone ($p < 0.01$; Appendix B). Peak magnitudes decrease to < 5 pieces cm^{-2} $peak^{-1}$ (Fig. 4). mFRIs (95% CI) at RP and XI are 223 (90-390) and 293 (225-360) yr, respectively (Table 2; Fig. 5). Because XI recorded only four FRIs (Table 3, Fig. 5), we deemed mFRIs between RP and XI sufficiently close to pool data from within this zone. The mFRI (95% CI) of the composite record is 251 (158-352) yr, significantly different from the composite record of the Shrub Tundra Zone ($p = 0.03$; Fig. 6; Appendix B).

Forest-tundra Zone, 8.5-5.5 (RP), 8.0-5.5 (XI), 7.5-5.5 (CO), 7.8-5.5 (WK) ka BP -- Decreases in *Populus* (< 10%) and increases in *Picea* (≥1%) pollen percentages mark the onset of the Forest-tundra Zone (Fig. 3; Appendix B). *Alnus* pollen percentages increase from trace amounts to > 50% starting around 7.25-7.50 ka BP (Fig 3; Appendix B), coinciding with the start of the CO and WK records (Appendix B). With the rise in *Alnus* pollen, SCD values decrease for comparisons to modern Boreal Forest and Forest Tundra (< 0.1), and probability-of-analog for these biomes increase to > 30-40% (Fig. 3; Appendix B). Discriminant analysis of *Picea* pollen at RP and WK (Appendix B) plus previous research (Brubaker et al. 1983) indicate the near exclusive presence of *P. glauca* in this zone.

CHARs at XI (median = 0.03 pieces cm^{-2} yr^{-1}) are higher than in all previous zones ($p < 0.01$), while CHARs at RP were intermediate between the Herb Tundra ($p < 0.01$) and Shrub Tundra zones ($p < 0.01$; medians = 0.02-0.03 pieces cm^{-2} yr^{-1}; Appendix B). Charcoal peak magnitude is similar to the pervious zone at RP, and generally remains below 5 pieces cm^{-2} $peak^{-1}$ at all sites (Fig. 4). mFRIs (95% CI) for RP, CO, and WK are 238 (158-324), 210 (145-277), and 131 (95-172) yr, respectively. Except for RP vs. WK ($p = 0.04$), FRI distributions do not differ among sites (Fig. 5; Appendix B). Given the statistical difference between RP and WK, the composite FRI distribution only includes FRIs from RP and CO. The composite records has a mFRI (95% CI) of 227 (173-282) yr, similar to the composite record from the Deciduous Woodland Zone, but significantly longer than the composite records from the Shrub Tundra Zone ($p = 0.02$; Fig. 6; Appendix B).

Boreal Forest Zone, 5.5 ka BP - present (RP, XI, CO, WK) -- *Picea* pollen percentages increase to > 10% at all sites between 6 and 4 ka BP (Fig. 3; Appendix B). With the rise in *Picea* pollen, all sites show an increase in the probability-of-analog with the modern Boreal Forest biome (> 75%) and lower probabilities for modern Forest-tundra (< ca 70%; Fig. 3; Appendix B). The first presence of *Picea* stomata ca 5.0 ka BP (RP; Fig. 3) and 5.4 ka BP (WK; Appendix B) coincides with the transition from Forest-tundra to Boreal Forest inferred from the modern analog analysis (Fig. 3; Appendix B). *Picea* pollen morphology from RP and WK (Appendix B) and previous research (Brubaker et al. 1983) indicate an increase of *P. mariana* at this time (Appendix B). While the beginning of this zone arguably differs across sites (e.g. RP vs. CO and

WK; Fig. 3, Appendix B), our results do not change when all sites are analyzed with a starting date of 5.0 vs. 5.5 ka BP (data not shown). We use 5.5 ka BP at all sites for simplicity.

CHARs increase to their highest levels in most records (median CHARs = 0.05-0.11 pieces cm^{-2} yr^{-1}; Fig. 4; Appendix B), although the period of increase varies by ca 500 to 1000 yr (CO, WK vs. RP; Fig. 4). mFRIs (95% CI) at RP, CO, and WK are 171 (135-216), 135 (113-160) and 135 (113-157) yr, respectively (Table 2; Fig. 5), and FRI distributions do not differ between sites (Fig. 5; Appendix B). The mFRI (95% CI) of the composite FRI distribution is 145 (129-163) yr; significantly shorter than mFRI in the composite record from the Forest-tundra and Deciduous Woodland Zones but similar to the composite record from the Shrub Tundra Zone (Fig. 6; Appendix B)[5].

DISCUSSION

Interpreting sediment charcoal records and detecting changes in fire regimes

We introduce three general tools that facilitate the interpretation of fire history from sediment-charcoal records. First, the signal-to-noise index provides a semi-objective way to judge if a record is appropriate for peak analysis. For example, while > 0.8 in most records, SNIs were consistently < 0.5 for the 8-0 ka BP in the Xindi Lake record (data not shown), indicating that this section was not suitable for peak identification. Second, our use of a Gaussian mixture model to determine threshold values for peak identification allowed us to treat all charcoal records with one set of objectively-determined criteria. These criteria are consistent with a

[5] To test the sensitivity our results to FRIs from WK, we constructed a pooled record for the Boreal Forest Zone that excluded WK. Significant differences remained when comparing the pooled Forest-tundra and Boreal Forest zones ($p = 0.047$), but not when comparing pooled Deciduous Woodland and Boreal Forest zones ($p = 0.072$).

mechanistic model of the origin of charcoal records (Higuera et al. 2007) and have the advantage of being established *a priori*. Further, because thresholds are defined locally, this approach is appropriate for records with changing variability in charcoal accumulation and is insensitive to variety of analytical choices (e.g. transforming charcoal data, defining C_{peak} via ratios vs. residuals; P.E. Higuera unpublished data). Our approach was generally successful, as the 99th-precentile criterion accurately identified known fires within 1 km of each lake, although a robust calibration requires a large dataset of known fires. Third, using the likelihood ratio test and pooling FRIs from multiple records greatly improved the ability to detect changes in fire regimes with long return intervals. In stand-replacing fire regimes with long and variable FRIs, individual records can detect only large or long-lasting changes in mFRIs (e.g. > 30-50% change over millennial time scales). Pooling data from several sites increased the sample size of FRIs and allowed the detection of statistical differences between pollen zones that was not possible using single records. This approach assumes that fire regimes are homogenous across sites and within time periods. We used between-site comparisons to evaluate the assumption of homogenous fire regimes, leading to the elimination of sites with statistically different records from the pooled data (i.e. WK within the Forest-tundra Zone)

Late-glacial and Holocene fire regimes: patterns and inferred controls

Herb Tundra Zone -- Though spanning a brief period of this zone, charcoal records suggest that fire was rare in the late-glacial herb tundra. Both climate and vegetation likely reduced the probability of fire. Summers were cooler and drier than present (Anderson and Brubaker 1994, Edwards et al. 2001, Anderson et al. 2004), with cold temperatures implying limited convection necessary for lightning ignition. The discontinuous herbaceous and low-shrub vegetation would have reduced fire spread if/when ignitions occurred. While species composition differed from modern tundra, the structure of vegetation in this zone may have been similar to present high-arctic tundra, where a cold and dry climate results in discontinuous vascular plant cover (Walker et al. 2005) that supports few fires (Kasischke et al. 2002).

Shrub Tundra Zone -- Fire activity increased markedly with the transition from herb to shrub tundra ca 13.3-14.3 ka BP, resulting in a mFRI (144 yr [119-170]; Fig. 6) statistically similar to those characterizing Alaskan boreal forests (Kasischke et al. 2002, Lynch et al. 2002; this study). This finding is surprising relative to modern tundra, since only 3% of Alaskan tundra burned between 1950 and 2004 (Alaska Fire Service 2004, Walker et al. 2005) and even the most flammable Alaskan tundra region, the Seward Peninsula (Fig. 1), has an estimated fire rotation period (analogous to a mFRI) of 270 yr (Kasischke et al. 2002).

Changes in both climate and vegetation favored frequent fires. Increased summer temperature from the Herb Tundra Zone and drier-than-present conditions (Anderson and Brubaker 1994, Abbott et al. 2000, Edwards et al. 2001, Anderson et al. 2004) would have favored fire ignition and spread. However, since temperatures remained cooler-than-present for most of the Shrub Tundra Zone, it is unlikely that temperature was the major driver of higher fire frequencies. The coincidence of increased CHARs and larger CHAR peaks with the expansion of shrub *Betula* (*B. glandulosa* and *B. nana*; Anderson and Brubaker 1994) suggests that the development of shrub tundra itself facilitated greater burning. *B. glandulosa*, with its small diameters and highly resinous stems (Dugle 1966), burns readily on modern landscapes (de Groot and Wein 2004), and high shrub densities would have provided the continuous flammable fuels necessary for fire spread. In addition, rapid revegetation following fire (de Groot and Wein

1999) would have facilitated short FRIs. Although FRIs were statistically similar to those in the Boreal Forest Zone (Fig. 5, 6), the lower overall CHARs and peak magnitudes (Fig. 4; Appendix B) in the Shrub Tundra Zone suggest that fuel loads were lower than in modern boreal forests.

Deciduous Woodland Zone – With the development of deciduous woodlands ca 10.5 ka BP, fires became less common (mFRI 251 yr [158-352], Fig. 6) and produced little charcoal (Fig. 4; Appendix B), implying that both climate and vegetation were less conducive to burning. However, given regional evidence that summers that were > 2° C warmer and 25-40% drier than present (Edwards et al. 2001, Anderson et al. 2004, Kaufman et al. 2004), one would predict an increase rather than decrease in fire activity during this period. Though inconsistent with direction of climatic change, a decline in fire would be expected given the lower flammability of deciduous vegetation. As in the modern boreal forest, deciduous trees in the past (both *Populus* and *Betula*; Edwards et al. 2005) would have acted as fire breaks, reducing fire spread across the landscape (Johnson 1992, Cumming 2001, Hely et al. 2001). Our findings contrast with those of Anderson et al. (2006), who infer that fires were common (mFRI of 77 yr, +/- 49 s.d.) when *Populus* pollen was abundant at Paradox Lake (Fig. 1), Kenai Peninsula, Alaska (8.5-10.7 ka BP). Although climatic difference between the Kenai Peninsula and Brooks Range may have resulted in different ignition rates and moisture levels, it is difficult to reconcile the short FRIs at Paradox Lake with the low flammability of the deciduous vegetation and the finding of low CHARs in this portion of the Paradox Lake record.

Forest-tundra Zone -- Fire return intervals decreased slightly, but not significantly, with the establishment of *P. glauca* in the mid-Holocene (mFRI 227 yr [173-282]; Fig. 6). Summer temperatures cooled and relative moisture increased in this zone (but remained drier than present; Abbott et al. 2000, Anderson et al. 2001, Edwards et al. 2001). Pollen and stomata data suggest that the vegetation resembled modern treeline, with *P. glauca* trees or stands dispersed within a landscape of *Betula* shrubs (Fig. 3; Appendix B). Although temperature and moisture trends would have reduced fire activity compared to the previous period, the unchanging mFRIs suggest that these climatic effects were balanced by the increase in landscape flammability resulting from the replacement of *Populus* with *P. glauca*. In addition increased CHARs and peak magnitudes in this zone (Fig. 4; Appendix B) suggest greater biomass burning per fire due to increased fuel loads. Unlike other zones, fire regimes varied across the study region, with significantly higher fire activity in the east (WK) compared to the west (RP; Fig. 5). Pollen records do not indicate a gradient in vegetation that would account for this pattern (Appendix B), and thus the shorter FRIs at WK suggest a gradient in climatic controls of fire during this period. Unfortunately, evaluating this possibility is difficult with existing paleoclimate records. Overall, mFRIs of this zone are at the lower end estimated fire rotation periods in modern forest-tundra (180-1000+ yr; Payette et al. 1989, Kasischke et al. 2002), possibly reflecting the generally warmer, drier climatic conditions during the Forest-tundra Zone compared to modern (Anderson et al. 2004).

Boreal Forest Zone – A decrease in mFRIs at two of the three study sites coincided with the development of *P. mariana*-dominated forests ca 5.5 ka BP (Table 2, Fig. 5), resulting in a pooled mFRI of 145 yr (129-163; Fig. 6). The absence of decreased FRIs at WK can be attributed to the significantly lower FRIs in the previous zone at this site (as described above). Peak magnitudes and/or CHARs reach their maxima at all sites (Fig. 4; Appendix B), suggesting greater charcoal production per fire, probably due to the abundant fine fuels of *P. mariana* and increased confer densities (Johnson 1992). Several lines of evidence indicate that effective moisture increased to near-modern levels by ca 5 ka BP and that temperatures continued to cool into the late Holocene (5-0 ka BP; Ellis and Calkin 1984, Evison et al. 1998, Abbott et al. 2000, Anderson et al. 2001, Anderson et al. 2004). Since these climatic changes should reduce fire ignition and spread, the increase in fire activity was likely due to increased landscape flammability associated with greater conifer density and the flammable fuels of *P. mariana*. This interpretation is consistent with several other Holocene fire-history studies from boreal Alaska (Dune, Low, Moose and Chokosna lakes, Fig. 1; Lynch et al. 2002, Lynch et al. 2004b, Hu et al. 2006) and with modeling studies showing higher fire frequencies with increased *P. mariana* abundance (Rupp et al. 2002). However, these findings contrast with the Paradox Lake record (Anderson et al. 2006), which indicates that FRIs increased from 81 (+/- 41 s.d.) to 130 (+/- 66 s.d.) yr with *P. mariana* arrival ca 4.6 ka BP. Thus although the FRIs for the boreal forest period are similar to the central Brooks Range, the direction of change differed. Perhaps the low density of *P. mariana* at Parodox Lake (Anderson et al. 2006) did not cause a large enough increase in landscape flammability to override the effects of cooler, moister climatic conditions on fire regimes.

Vegetation mediates the impacts of climatic change on fire regimes

Despite numerous examples that millennial-scale climatic changes have driven variations in Holocene fire regimes (e.g. Clark 1988, Millspaugh et al. 2000, Carcaillet et al. 2001, Gavin et al. 2006), paleoecological records from central Brooks Range indicate that climatic change was not the primary cause of many fire regime shifts. A major theme of fire-history records from this region is that fuel characteristics and abundance amplified or dampened the direct effects of millennial-scale climatic changes (Hu et al. 2006). For example, in response to climatic warming during late glaciation and the early Holocene, vegetation shifted from herb tundra to *Betula*-dominated shrub tundra (Anderson et al. 2004; Fig. 3). While warming likely facilitated increased burning, the evidence of cooler-than-present summers implies that fuels were of greater influence in creating fire frequencies similar to those in modern boreal forests (Fig. 6).

A more unexpected outcome of vegetation change occurs when vegetation negates the direct effect of climate by changing the probability of fire in the opposite direction of climatic change. We infer two examples of this effect. First, in the mid-to-late Holocene, a cooler and wetter climate reduced the probability of fire, but an increase of *P. mariana* and overall conifer densities increased vegetation flammability. The net result of these shifts in vegetation and climate was greater fire activity (Fig. 5-6), with the positive effect of fuels overshadowing the negative effect of climate on fire occurrence. Second, despite warmer-than-present summers during the early Holocene, fire activity decreased when deciduous woodlands expanded into shrub tundra (Fig. 6). In this case, the low flammability of deciduous trees had a greater influence than increased temperatures on the probability of fire. Overall, the postglacial fire history of the southcentral Brooks Range emphasizes the central role of fuels to past and, by inference, to future fire regimes.

Implications for global change in arctic and subarctic ecosystems

Fire regimes in past herb tundra, shrub tundra, and deciduous woodlands reflect the effects of climates and vegetation biomes that do not have counterparts on the modern landscape (i.e. no analog climate and vegetation, Anderson et al. 1989, Bartlein et al. 1991, Williams and Jackson 2007, Appendix B). Given the potential for novel vegetation and climate in the future (Edwards et al. 2005, Williams and Jackson 2007), our results have implications for anticipating future fire regimes in arctic and subarctic ecosystems.

Our finding that vegetation mediated the impacts of climate change emphasizes the importance of biological-physical feedbacks in past arctic and subarctic ecosystems. While direct linkages between climate and fire may predict future fire regimes at annual to decadal time scales (Kasischke et al. 2002, Lynch et al. 2004b, Duffy et al. 2005), the paleo-record highlights that feedbacks between climate, vegetation, and fire can override direct climatic effects at multi-decadal to centennial time scales. For example, in boreal forests the direct effect of future climatic warming should be to increase area burned, with a secondary effect of replacing coniferous with deciduous forest types (Rupp et al. 2000, Calef et al. 2005, Flannigan et al. 2005, Johnstone and Chapin 2006). Our finding of few fires during the warmer-than-present Deciduous Woodland Zone implies that future increases in burning could lower the probability of subsequent fires by favoring succesional forests with less flammable, deciduous fuels. If deciduous stands are maintained across the landscape via gap-phase replacement (Cumming et al. 2000, Johnstone and Chapin 2006), this negative feedback mechanism could result in fires being less frequent than would be predicted by climate’s direct effect on area burned.

The importance of biological-physical feedbacks is also highlighted by high fire frequencies in past shrub tundra. Our records provide a clear precedence that shrub-dominated tundra can sustain higher fire frequencies than present-day tundra. Thus, the future expansion of tundra shrubs (Tape et al. 2006, Walker et al. 2006) coupled with decreased effective moisture

(ACIA 2004) could enhance circumarctic burning and initiate feedbacks with the climate system. Recent studies of modern tundra fires suggest the possibility for both short and long term impacts of increased tundra burning ranging from increased summer soil temperatures and moisture levels (Liljedahl et al. 2007) to the release ancient soil carbon from increased permafrost thawing and organic-layer consumption (Racine et al. 2006, Liljedahl et al. 2007). Given the concern over the fate of terrestrial carbon in tundra and other high-latitude ecosystems (Zimov et al. 1999, Chapin et al. 2000, Mack et al. 2004, Weintraub and Schimel 2005), the evidence of fires in early Holocene tundra should motivate research into the controls of tundra fire regimes and linkages between tundra burning and the climate system.

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LITERATURE CITED

Abbott, M. B., B. P. Finney, M. E. Edwards, and K. R. Kelts. 2000. Lake-level reconstruction and paleohydrology of Birch Lake, central Alaska, based on seismic reflection profiles and core transects. Quaternary Research **53**:154-166.

ACIA. 2004. Impacts of a Warming Arctic: Arctic Climate Impact Assessment. Cambridge University Press, Cambridge.

Alaska Fire Service. 2004. Alaska Fire History. Bureau of Land Management. Bureau of Land Management. Alaska Fire Service.

Anderson, L., M. B. Abbott, and P. B. Finney. 2001. Holocene climate inferred from oxygen isotope ratios in lake sediments, Central Brooks Range, Alaska. Quaternary Research **55**:313-321.

Anderson, P. M., P. J. Bartlein, L. B. Brubaker, K. Gajewski, and J. C. Ritchie. 1989. Modern analogs of late-Quaternary pollen spectra from the western interior of North America. Journal of Biogeography **16**:573-596.

Anderson, P. M. and L. B. Brubaker. 1994. Vegetation history of northcentral Alaska: a mapped summary of Late-Quaternary pollen data. Quaternary Science Reviews **13**:71-92.

Anderson, P. M., M. E. Edwards, and L. B. Brubaker. 2004. Results and paleoclimate implications of 35 years of paleoecological research in Alaska. Developments in Quaternary Science **1**:427-440.

Anderson, R. S., D. J. Hallett, E. Berg, R. B. Jass, J. L. Toney, C. S. de Fontaine, and A. DeVolder. 2006. Holocene development of Boreal forests and fire regimes on the Kenai Lowlands of Alaska. Holocene **16**:791-803.

Bartlein, P. J., P. M. Anderson, M. E. Edwards, and P. F. McDowell. 1991. A framework for interpreting paleoclimatic variations in eastern Beringia. Quaternary International **10-12**:73-83.

Binford, M. W. 1990. Calculation and uncertainty analysis of 210Pb dates for PIRLA project lake sediment cores. Journal of Paleolimnology **3**:253-267.

Brubaker, L. B., H. L. Garfinkel, and M. E. Edwards. 1983. A late Wisconsin and Holocene vegetation history from the central Brooks Range: Implications for Alaskan paleoecology. Quaternary Research **20**:194-214.

Calef, M. P., A. D. McGuire, H. E. Epstein, T. S. Rupp, and H. H. Shugart. 2005. Analysis of vegetation distribution in interior Alaska and sensitivity to climate change using a logistic regression approach. Journal of Biogeography **32**:863-878.

Carcaillet, C., Y. Bergeron, P. J. H. Richard, B. Frechette, S. Gauthier, and Y. T. Prairie. 2001. Change of fire frequency in the eastern Canadian boreal forests during the Holocene: does vegetation composition or climate trigger the fire regime? Journal of Ecology **89**:930-946.

Carlson, L. J. 2003. Describing the postglacial pattern and rate of *Picea* expansion in Alaska using paleoecological records. Ph.D. dissertation. University of Washington, Seattle.

Chapin, F. S., A. D. McGuire, J. Randerson, R. S. Pielke, D. Baldocchi, S. E. Hobbie, N. Roulet, W. Eugster, E. Kasischke, E. B. Rastetter, S. A. Zimov, and S. W. Running. 2000. Arctic and boreal ecosystems of western North America as components of the climate system. Global Change Biology **6**:211-223.

Clark, J. S. 1988. Effects of climate change on fire regimes in northwestern Minnesota. Nature **334**:233-235.

Clark, J. S. 1989. Ecological disturbance as a renewal process: theory and application to fire history. Oikos **56**:17-30.

Clark, J. S. 1990. Fire and climate change during the last 750 yr in northwestern Minnesota. Ecological Monographs **60**:135-159.

Clark, J. S., P. D. Royall, and C. Chumbley. 1996. The role of fire during climate change in an eastern deciduous forest at Devil's Bathtub, New York. Ecology **77**:2148-2166.

Cumming, S. G. 2001. Forest type and wildfire in the Alberta boreal mixedwood: What do fires burn? Ecological Applications **11**:97-110.

Cumming, S. G., F. K. A. Schmiegelow, and P. J. Burton. 2000. Gap dynamics in boreal aspen stands: Is the forest older than we think? Ecological Applications **10**:744-759.

de Groot, W. J. and R. W. Wein. 1999. *Betula glandulosa* Michx. response to burning and postfire growth temperature and implications of climate change. International Journal of Wildland Fire **9**:51-64.

de Groot, W. J. and R. W. Wein. 2004. Effects of fire severity and season of burn on *Betula glandulosa* growth dynamics. International Journal of Wildland Fire **13**:287-295.

Duffy, P. A., J. E. Walsh, J. M. Graham, D. H. Mann, and T. S. Rupp. 2005. Impacts of large-scale atmospheric-ocean variability on Alaskan fire season severity. Ecological Applications **15**:1317-1330.

Dugle, J. R. 1966. A taxonomic study of Western Canadian species in genus *Betula*. Canadian Journal of Botany **44**:929-&.

Edwards, M. E., L. B. Brubaker, A. V. Lozhkin, and P. M. Anderson. 2005. Structurally novel biomes: a response to past warming in Beringia. Ecology **86**:1696-1703.

Edwards, M. E., C. J. Mock, B. P. Finney, V. A. Barber, and P. J. Bartlein. 2001. Potential analogues for paleoclimatic variations in eastern interior Alaska during the past 14,000 yr: atmospheric-circulation controls of regional temperature and moisture responses. Quaternary Science Reviews **20**:189-202.

Ellis, J. M. and P. E. Calkin. 1984. Chronology of Holocene glaciation, central Brooks Range, Alaska. Geological Society of America Bulletin **95**:897-912.

Evison, L. H., P. E. Calkin, and J. M. Ellis. 1998. Late-Holocene glaciation and twentieth-century retreat, northeastern Brooks Range, Alaska. The Holocene **6**:17-24.

Flannigan, M. D., K. A. Logan, B. D. Amiro, W. R. Skinner, and B. J. Stocks. 2005. Future area burned in Canada. Climatic Change **72**:1-16.

Gavin, D. G., F. S. Hu, K. Lertzman, and P. Corbett. 2006. Weak climatic control of stand-scale fire history during the late Holocene. Ecology **87**:1722-1732.

Hansen, B. C. S. 1994. Conifer stomate analysis as a paleoecological tool: an example from the Hudson Bay Lowlands. Canadian Journal of Botany **73**:244-252.

Hely, C., M. Flannigan, Y. Bergeron, and D. McRae. 2001. Role of vegetation and weather on fire behavior in the Canadian mixedwood boreal forest using two fire behavior prediction systems. Canadian Journal of Forest Research **31**:430-441.

Higuera, P. E., M. E. Peters, L. B. Brubaker, and D. G. Gavin. 2007. Understanding the origin and analysis of sediment-charcoal records with a simulation model. Quaternary Science Reviews **26**:1790-1809.

Higuera, P. E., D. G. Sprugel, and L. B. Brubaker. 2005. Reconstructing fire regimes with charcoal from small-hollow sediments: a calibration with tree-ring records of fire. The Holocene **15**:238-251.

Hu, F. S., L. B. Brubaker, D. G. Gavin, P. E. Higuera, J. A. Lynch, T. S. Rupp, and W. Tinner. 2006. How climate and vegetation influence the fire regime of the Alaskan Boreal Biome: the Holocene perspective. Mitigation and Adaptation Strategies for Global Change **11**:829-846.

Johnson, E. A. 1992. Fire and Vegetation Dynamics: Studies from the North American Boreal Forest. Cambridge University Press, Cambridge.

Johnson, E. A. and S. L. Gutsell. 1994. Fire frequency models, methods and interpretations. Advances in Ecological Research **25**:239-287.

Johnstone, J. F. and F. S. Chapin. 2006. Fire interval effects on successional trajectory in boreal forests of northwest Canada. Ecosystems **9**:268-277.

Kasischke, E. S. and M. R. Turetsky. 2006. Recent changes in the fire regime across the North American boreal region - Spatial and temporal patterns of burning across Canada and Alaska. Geophysical Research Letters **33**.

Kasischke, E. S., D. Williams, and D. Barry. 2002. Analysis of the patterns of large fires in the boreal forest region of Alaska. International Journal of Wildland Fire **11**:131-144.

Kaufman, D. S., T. A. Ager, N. J. Anderson, P. M. Anderson, J. T. Andrews, P. J. Bartlein, L. B. Brubaker, L. L. Coats, L. C. Cwynar, M. L. Duvall, A. S. Dyke, M. E. Edwards, W. R. Eisner, K. Gajewski, A. Geirsdottir, F. S. Hu, A. E. Jennings, M. R. Kaplan, M. W. Kerwin, A. V. Lozhkin, G. M. MacDonald, G. H. Miller, C. J. Mock, W. W. Oswald, B. L. Otto-Bliesner, D. F. Porinchu, K. Ruhland, J. P. Smol, E. J. Steig, and B. B. Wolfe. 2004. Holocene thermal maximum in the western Arctic (0-180 degrees W). Quaternary Science Reviews **23**:529-560.

Liljedahl, A., L. Hinzman, R. Busey, and K. Yoshikawa. 2007. Physical short-term changes after a tussock tundra fire, Seward Peninsula, Alaska. Journal of Geophysical Research-Earth Surface **112**.

Lloyd, A. H. 2005. Ecological histories from Alaskan tree lines provide insight into future change. Ecology **86**:1687-1695.

Lynch, J. A., J. S. Clark, N. H. Bigelow, M. E. Edwards, and B. P. Finney. 2002. Geographic and temporal variations in fire history in boreal ecosystems of Alaska. Journal of Geophysical Research **108**:FFR8-1-FFR8-17.

Lynch, J. A., J. S. Clark, and B. J. Stocks. 2004a. Charcoal production, dispersal and deposition from the Fort Providence experimental fire: Interpreting fire regimes from charcoal records in boreal forests. Canadian Journal of Forest Research **34**:1642-1656.

Lynch, J. A., J. L. Hollis, and F. S. Hu. 2004b. Climatic and landscape controls of the boreal forest fire regime: Holocene records from Alaska. Journal of Ecology **92**:447-489.

Mack, M. C., E. A. G. Schuur, M. S. Bret-Harte, G. R. Shaver, and F. S. Chapin. 2004. Ecosystem carbon storage in arctic tundra reduced by long-term nutrient fertilization. Nature **431**:440-443.

Millspaugh, S. H., C. Whitlock, and P. Bartlein. 2000. Variations in fire frequency and climate over the past 17000 yr in central Yellowstone National Park. Geology **28**:211-214.

Nowacki, G., P. Spencer, T. Brock, M. Fleming, and T. Jorgenson. 2000. Narrative Description for the Ecoregions of Alaska and Neighboring Territories. USGS, Reston, VA.

Overpeck, J., K. Hughen, D. Hardy, R. Bradley, R. Case, M. Douglas, B. Finney, K. Gajewski, G. Jacoby, A. Jennings, S. Lamourex, A. Lasca, G. MacDonald, J. Moore, M. Retelle, S.

Smith, A. Wolfe, and G. Zielinski. 1997. Arctic environmental change of the last four centuries. Science **278**:1251-1256.

PALE members. 1994. Research Protocols for PALE: Paleoclimates of Arctic Lakes and Estuaries. PAGES workshop report:53.

Payette, S., C. Morneau, L. Sirois, and M. Desponts. 1989. Recent fire history of the northern Quebec biomes. Ecology **70**:656-673.

Racine, C., J. A. Allen, and J. G. Dennis. 2006. Long-term monitoring of vegetation change following tundra fires in Noatak National preserve, Alaska. NPS/AKRARCN/NRTR-2006/02

Arctic Network of Parks Inventory and Monitoring Program, National Park Service, Alaska Region.

Rupp, T. S., F. S. I. Chapin, and A. M. Starfield. 2000. Response of subarctic vegetation to transient climatic change on the Seward Peninsula in northwest Alaska. Global Change Biology **6**:541-555.

Rupp, T. S., A. M. Starfield, F. S. I. Chapin, and P. Duffy. 2002. Modeling the impact of black spruce on the fire regime of Alaskan boreal forest. Climatic Change **55**:213-233.

Serreze, M. C., J. E. Walsh, F. S. I. Chapin, T. Osterkamp, M. Dyurgerov, V. Romanovsky, W. C. Oechel, J. Morison, T. Zhang, and R. G. Barry. 2000. Observational evidence of recent change in the northern high-latitude environment. Climatic Change **46**:159-207.

Soja, A. J., N. M. Tchebakova, N. H. F. French, M. D. Flannigan, H. H. Shugart, B. J. Stocks, A. I. Sukhinin, E. I. Varfenova, F. S. Chapin, and P. W. Stackhouse. 2007. Climate-induced boreal forest change: Predictions versus current observations. Global and Planetary Change

56:274-296.

Tape, K., M. Sturm, and C. Racine. 2006. The evidence for shrub expansion in Northern Alaska and the Pan-Arctic. Global Change Biology **12**:686-702.

Telford, R. J., E. Heegaard, and H. J. B. Birks. 2004. All age-depth models are wrong: but how badly? Quaternary Science Reviews **23**:1-5.

Walker, D. A., M. K. Raynolds, F. J. A. Daniels, E. Einarsson, A. Evlvebakk, W. A. Could, A. E. Katenin, S. S. Kholod, C. J. Markon, E. S. Melnikov, N. G. Moskalenko, S. S. Talbot, B. A. Yurtsev, and o. m. o. t. C. Team. 2005. The circumpolar Arctic vegetation map. Journal of Vegetation Science **16**:267-282.

Walker, M. D., C. H. Wahren, R. D. Hollister, G. H. R. Henry, L. E. Ahlquist, J. M. Alatalo, M. S. Bret-Harte, M. P. Calef, T. V. Callaghan, A. B. Carroll, H. E. Epstein, I. S. Jonsdottir, J. A. Klein, B. Magnusson, U. Molau, S. F. Oberbauer, S. P. Rewa, C. H. Robinson, G. R. Shaver, K. N. Suding, C. C. Thompson, A. Tolvanen, O. Totland, P. L. Turner, C. E. Tweedie, P. J. Webber, and P. A. Wookey. 2006. Plant community responses to experimental warming across the tundra biome. Proceedings of the National Academy of Sciences of the United States of America **103**:1342-1346.

Weintraub, M. N. and J. P. Schimel. 2005. Nitrogen cycling and the spread of shrubs control changes in the carbon balance of arctic tundra ecosystems. Bioscience **55**:408-415.

Whitlock, C., M. M. Bianchi, P. J. Bartlein, V. Markgraf, J. Marlon, M. Walsh, and N. McCoy. 2006. Postglacial vegetation, climate, and fire history along the east side of the Andes (lat 41-42.5 degrees S), Argentina. Quaternary Research **66**:187-201.

Williams, J. W. and S. T. Jackson. 2007. Novel climates, no-analog communities, and ecological surprises. Frontiers in Ecology and the Environment **5**:475-482.

Wright, H. E., D. H. Mann, and P. H. Glaser. 1984. Piston corers for peat and lake sediments. Ecology **65**:657-659.

Yarie, J. 1981. Forest fire cycles and life tables: a case study from interior Alaska. Canadian Journal of Forest Research **11**:554-562.

Zar, J. H. 1999. Biostatistical Analysis. Fourth edition. Prentice Hall, Upper Saddle River.

Zimov, S. A., S. P. Davidov, G. M. Zimova, A. I. Davidova, F. S. Chapin, M. C. Chapin, and J. F. Reynolds. 1999. Contribution of disturbance to increasing seasonal amplitude of atmospheric CO2. Science **284**:1973-1976.

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TABLES

Table 1. Lake locations, characteristics, and record quality.

Lake Name (unofficial)	N Lattitude	W Longitude	Elevation (m asl)	Surface area (ha)	Depth at corring Site (m)	Age of record (ka)	Mean sedimentation rate +/- s.d. (cm yr^{-1})	Mean resolution +/- s.d. (yr sample^{-1})	Median signal-to-noise index
Ruppert	67°04’16”	154°14’45”	230	3	7.0	14.0	0.040 ± 0.023	13 ± 6	0.88
Xindi	67°06’42”	152°29’30”	240	7	10.6	15.5	0.025 ± 0.016	32† ± 24	0.71†
Code	67°09’29”	151°51’40”	250	2	7.0	7.5	0.017 ± 0.003	16 ± 3	0.83
Wild Tussock	67°07’40”	151°22’55”	290	15	11.6	8.0	0.019 ± 0.001	14 ± 5	0.84

† mean (s.d.) sample resolution and the signal-to-noise index for the section of core used for charocal peak identification was 23 (13) yrs, and 0.83, respectivly.

Table 2. Fire regime statistics for each site, stratified by pollen-defined vegetation zone. Parentheses enclose 95% confidence intervals estimated by 1000 bootstrapped samples of the FRI distributions. "N_{FRI}" is the number of fire return intervals in each zone (the total number of fires in a zone, minus one).

			Fire-history parameter (95% CI)		
Site	Zone	N_{FRI}	Mean fire return interval (yr)	Weibull b parameter (yr)	Weibull c parameter (unitless)
Ruppert	Shrub Tundra	20	137 (107 - 171)	151 (116 - 191)	1.84 (1.43 - 3.35)
Ruppert	Deciduous Woodland	6	223 (90 - 390)	229 (94 - 429)	1.16 (0.90 - 3.11)
Ruppert	Forest-tundra	12	238 (158 - 324)	262 (172 - 360)	1.63 (1.30 - 2.75)
Ruppert	Boreal Forest	31	171 (135 - 216)	188 (147 - 239)	1.53 (1.31 - 2.06)
Xindi	Shrub Tundra	24	150 (113 - 189)	164 (122 - 207)	1.68 (1.45 - 2.25)
Xindi	Deciduous Woodland	4	293 (225 - 360)	Weibull model not fit (< 5 FRIs)	
Code	Forest-tundra	8	210 (145 - 277)	235 (162 - 302)	2.39 (1.78 - 5.2)
Code	Boreal Forest	39	135 (113 - 160)	150 (123 - 178)	1.85 (1.52 - 2.60)
Wild Tussock	Forest-tundra	16	131 (95 - 172)	145 (104 - 191)	1.66 (1.38 - 2.5)
Wild Tussock	Boreal Forest	39	135 (113 - 157)	149 (123 - 174)	1.96 (1.61 - 2.75)

FIGURE LEGENDS

Figure 1. Location of lakes in this study and others discussed in the text (1, Dune Lake; 2, Low Lake; 3, Moose Lake; 4, Chokosna Lake; 5, Paradox Lake). Grey polygons are areas that have burned between CE 1950-2003 (Alaska Fire Service 2004), and the dashed line on the lower map is the southern border of Gates of the Arctic National Park. The black dots and larger circles identifying each lake on the bottom map have 1 and 2 km radii, representing the approximate spatial scale of each fire history record.

Figure 2. Age-depth models for each site with the resulting sedimentation rates and sample resolution, with 95% confidence intervals. At RP, temporal resolution changes from ca 25 to 10 yr sample^{-1} at 2.2 ka BP because sampling intervals change from 0.5 to 0.25 cm. *see Results for explanation.

Figure 3. Pollen and spore percentages of selected taxa; total pollen accumulation rate (PAR); squared chord distance (SCD) and probability of analog values for comparisons between fossil samples and those from modern Boreal Forest, Forest-tundra, and Arctic Tundra vegetation zones; and charcoal accumulation rate (CHAR) for Ruppert Lake. Filled (empty) circles on *Picea* panel represent *Picea* stomata presence (absence). Triangles below lower horizontal axis represent the location of ^{14}C or ^{210}Pb dates. See Appendix B for same figures for Xindi, Code, and Wild Tussock lakes.

Figure 4. (i) $C_{interpolated}$ (black) and $C_{background}$ (grey) series; (ii) C_{peak} series with the values identifying noise-related variability (positive and negative grey lines) and peaks identified with each threshold criterion. The 99th percentile criterion used for interpretation is represented with '+', and the 95th and 99.9th percentile results are represented with '**.**'. (iii) Pollen-inferred

vegetation zone and peak magnitude for all CHAR values exceeding the positive threshold value in (ii). Note, peak magnitude values not corresponding to “+” symbols in (ii) are those that did not pass the minimum-count screening (see Methods for details), and “+” symbols in (ii) with no apparent peak magnitude value correspond to very small peak magnitudes.

Figure 5. Distribution of FRIs and fitted Weibull models for each vegetation zone (columns) at each site (rows). Results from statistical comparisons are summarized by “=”, not different ($p > 0.05$), or “=/=”, different ($p \leq 0.05$) . Appendix B contains p-values for all comparisons.

Figure 6. Results from the analysis of pooled FRIs: distributions of FRIs, fitted Weibull models, Weibull *b* and *c* parameters (95% CI), mean fire return intervals (mFRI; 95% CI), and number of FRIs in each vegetation zone. Results from statistical comparisons are summarized as in Fig. 5 and presented in Appendix B. NOTE: Wild Tussock Lake (WK) is not included in the pooled record for the Forest-tundra Zone because of statistical differences between WK and Ruppert Lake (RP) during this period (Fig. 5; Appendix B).

Figure 1.

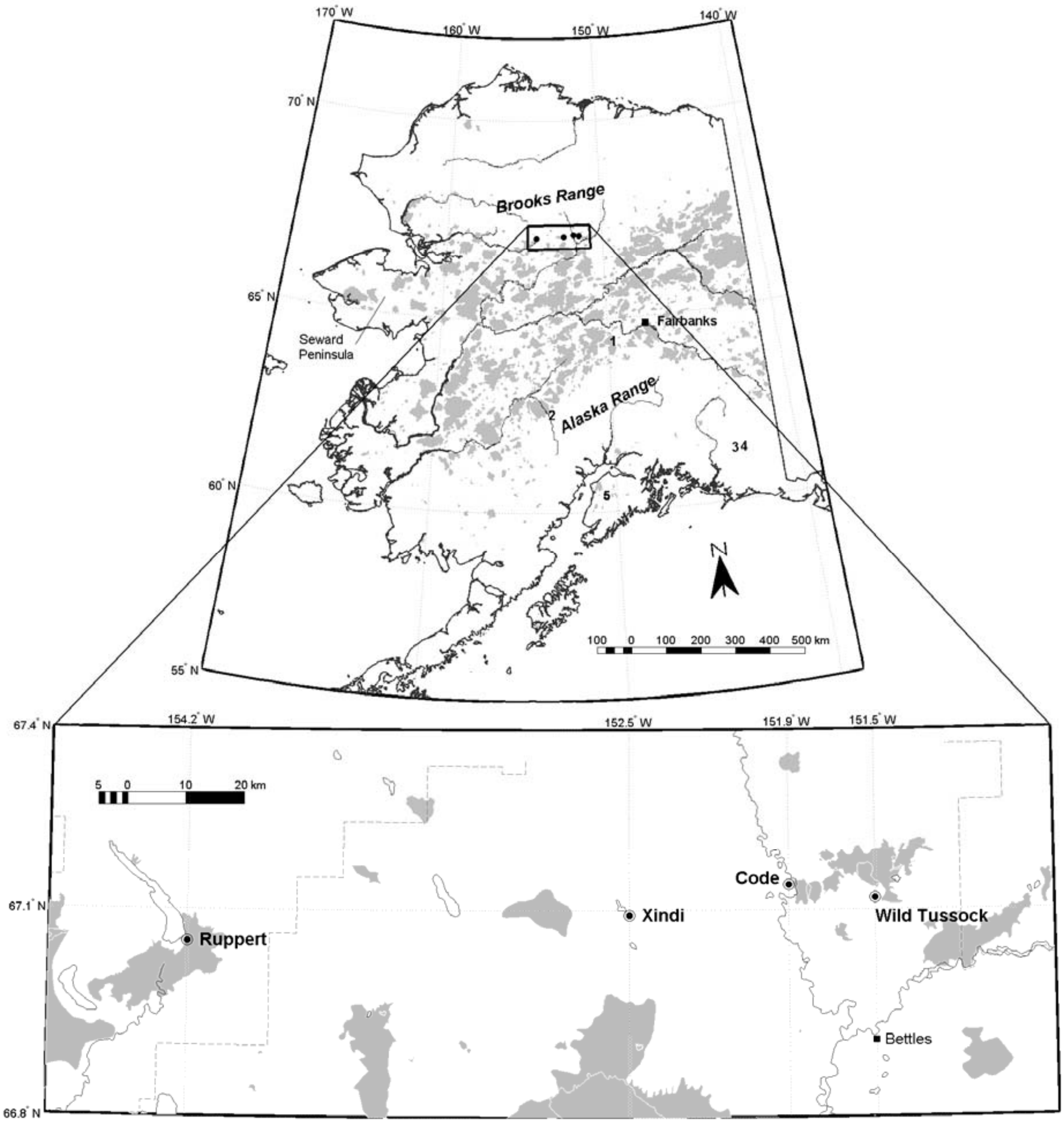

170° W
160° W
150° W
140° W
70° N
65° N
60° N
55° N
Brooks Range
Seward
Peninsula
Fairbanks
1
2
Alaska Range
34
5
N
100 0 100 200 300 400 500 km
67.4° N
67.1° N
66.8° N
154.2° W
152.5° W
151.9° W
151.5° W
5 0 10 20 km
Ruppert
Xindi
Code
Wild Tussock
Bettles

Figure 2.

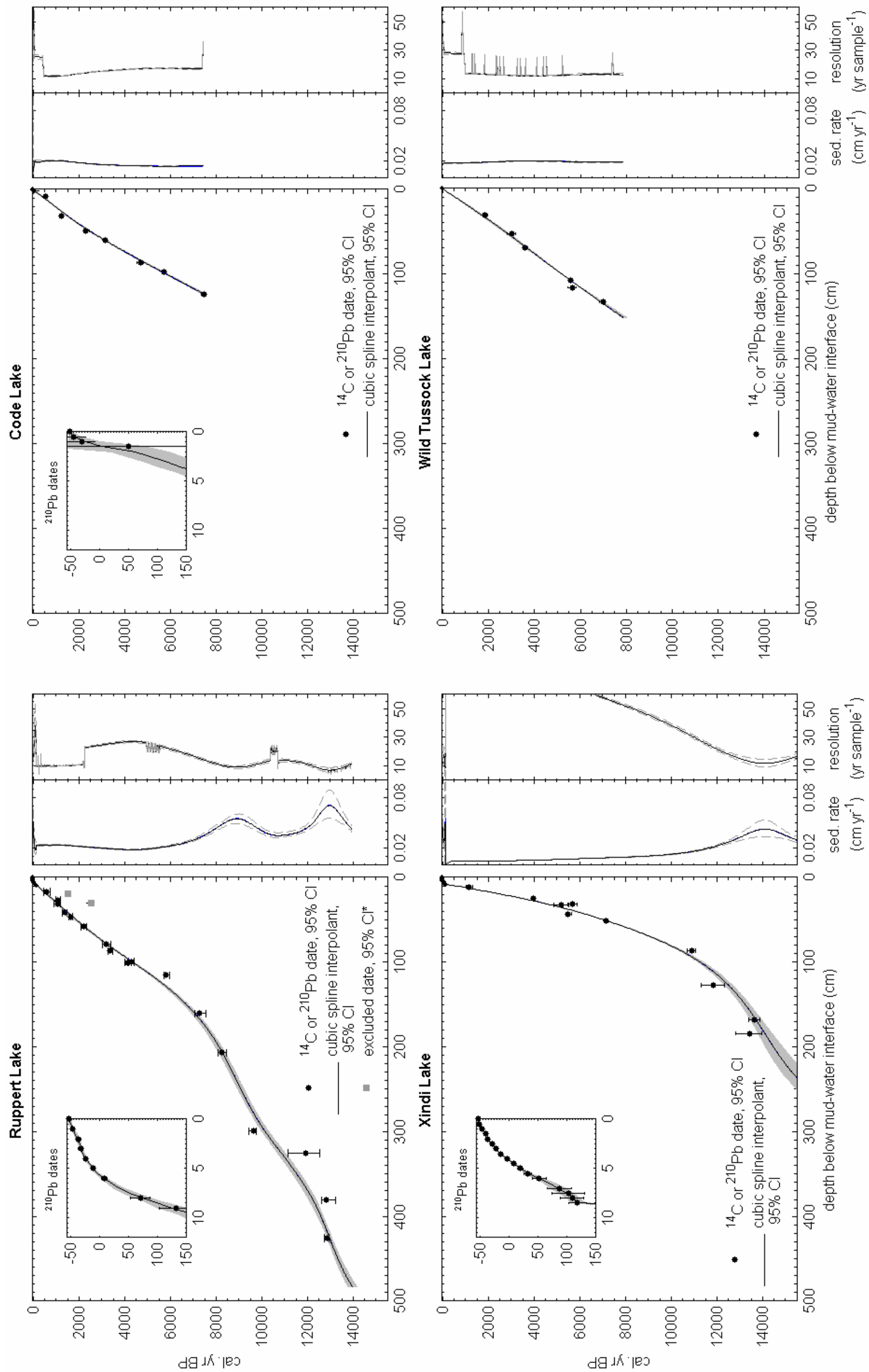
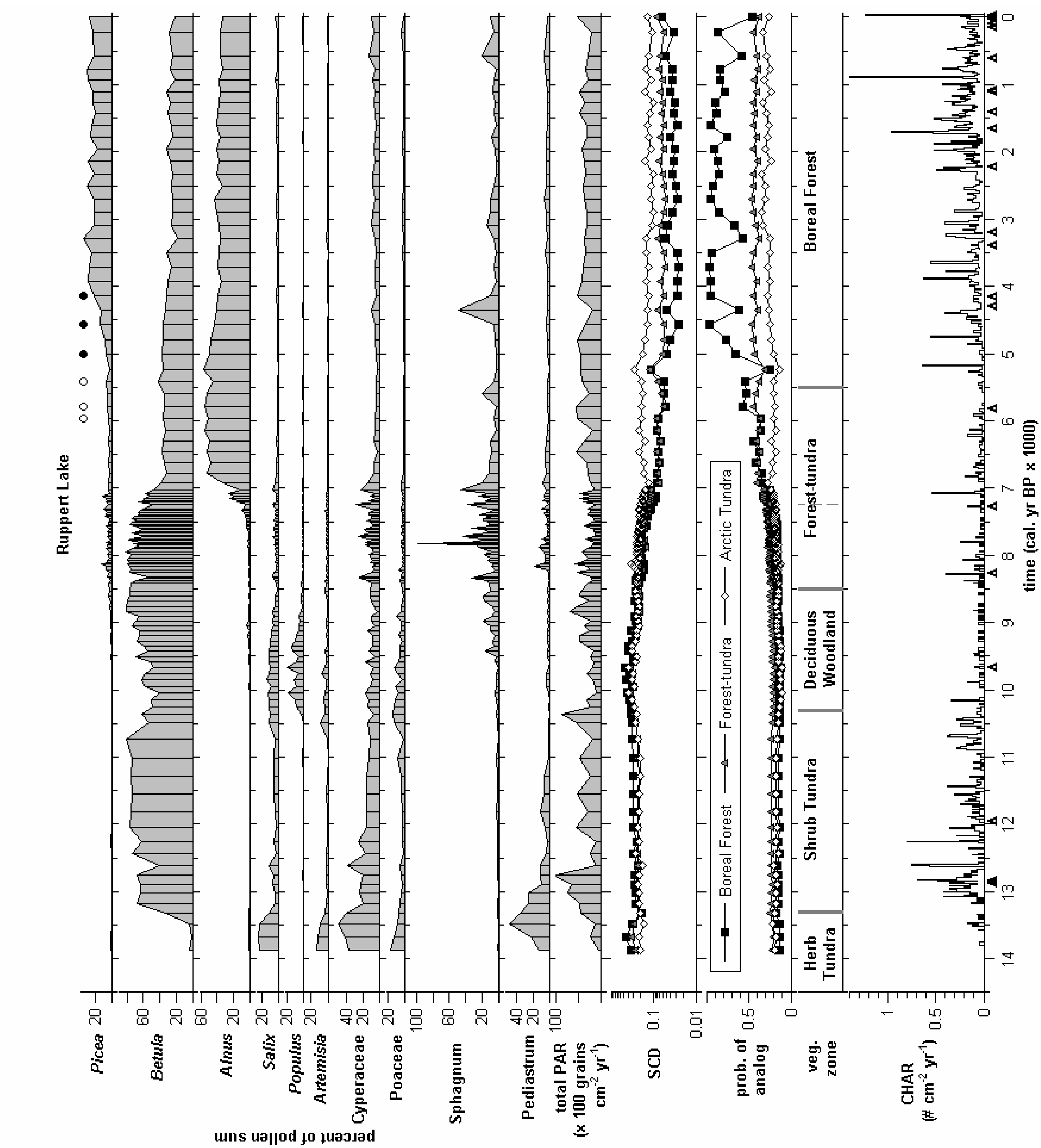
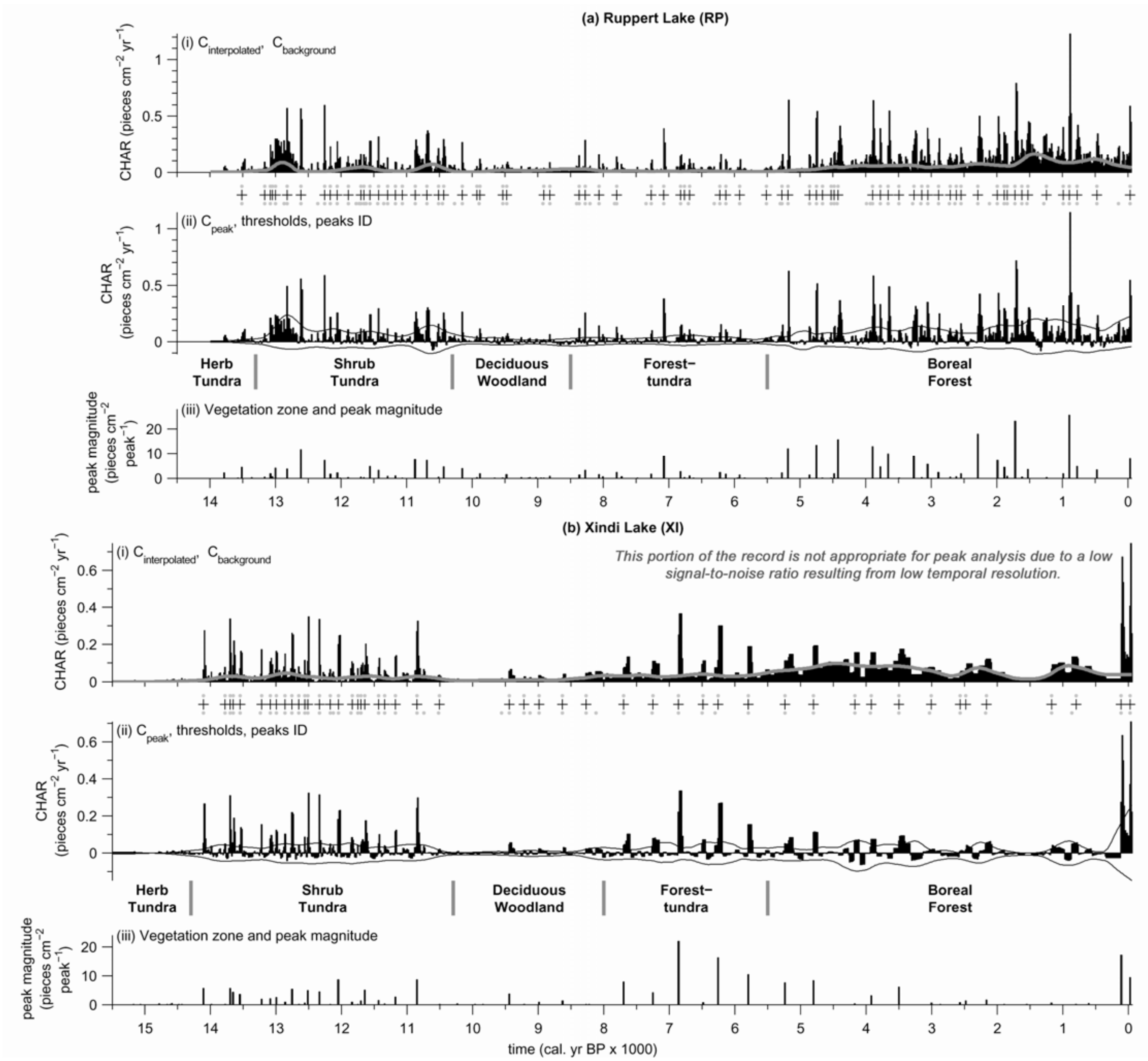

Ruppert Lake
Code Lake
Xindi Lake
Wild Tussock Lake
^{210}Pb dates
cal. yr BP
depth below mud-water interface (cm)
sed. rate (cm yr^{-1})
resolution (yr sample^{-1})
^{14}C or ^{210}Pb date, 95% CI
cubic spline interpolant, 95% CI
excluded date, 95% CI*

Figure 3.


Ruppert Lake
Picea
Betula
Alnus
Salix
Populus
Artemisia
Cyperaceae
Poaceae
Sphagnum
Pediastrum
total PAR (x 100 grains cm^-2 yr^-1)
percent of pollen sum
SCD
prob. of analog
veg. zone
CHAR (# cm^-2 yr^-1)
Boreal Forest
Forest-tundra
Arctic Tundra
Herb Tundra
Shrub Tundra
Deciduous Woodland
time (cal. yr BP x 1000)

1 Figure 4.


(a) Ruppert Lake (RP)
(i) C_interpolated, C_background
CHAR (pieces cm−2 yr−1)
(ii) C_peak, thresholds, peaks ID
CHAR (pieces cm−2 yr−1)
Herb Tundra
Shrub Tundra
Deciduous Woodland
Forest−tundra
Boreal Forest
(iii) Vegetation zone and peak magnitude
peak magnitude (pieces cm−2 peak−1)
(b) Xindi Lake (XI)
(i) C_interpolated, C_background
This portion of the record is not appropriate for peak analysis due to a low signal-to-noise ratio resulting from low temporal resolution.
CHAR (pieces cm−2 yr−1)
(ii) C_peak, thresholds, peaks ID
CHAR (pieces cm−2 yr−1)
Herb Tundra
Shrub Tundra
Deciduous Woodland
Forest−tundra
Boreal Forest
(iii) Vegetation zone and peak magnitude
peak magnitude (pieces cm−2 peak−1)
time (cal. yr BP x 1000)

2

Figure 4 – continued.

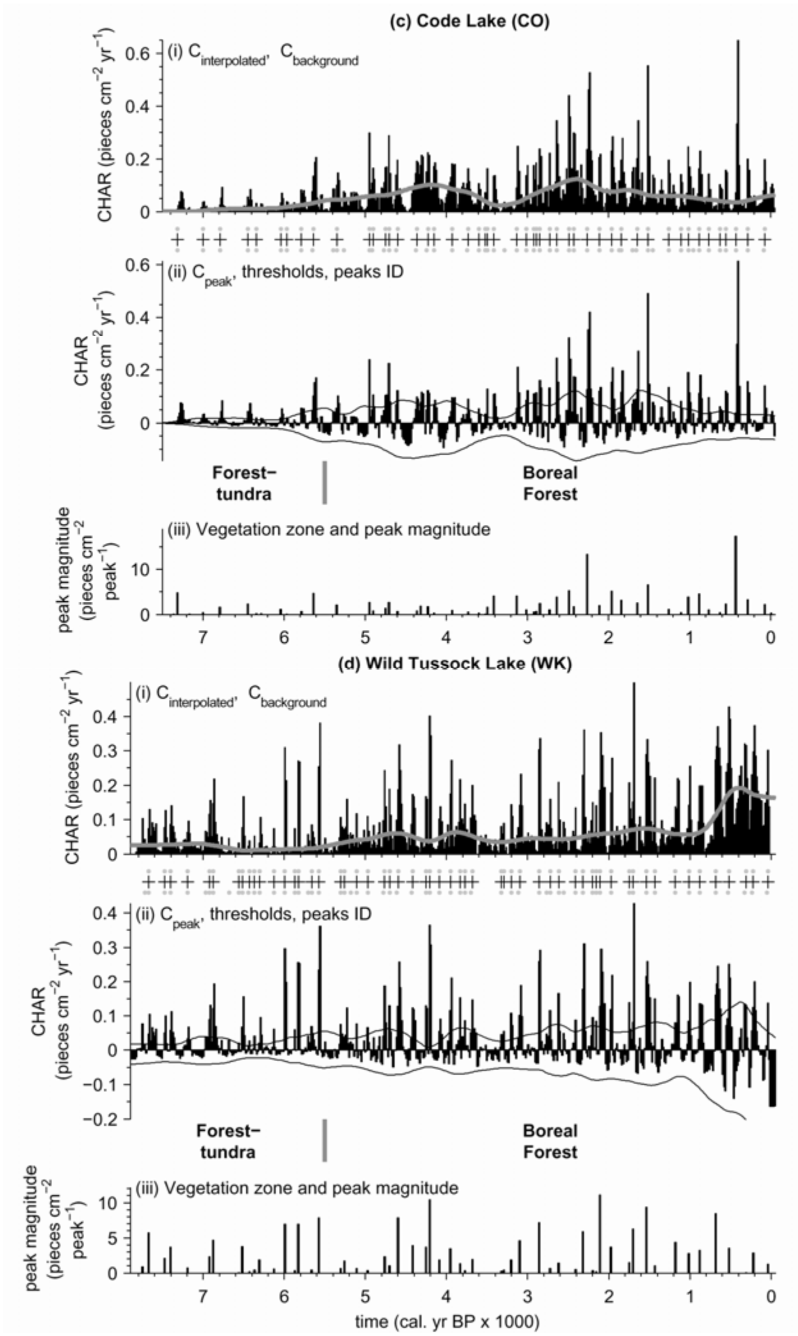
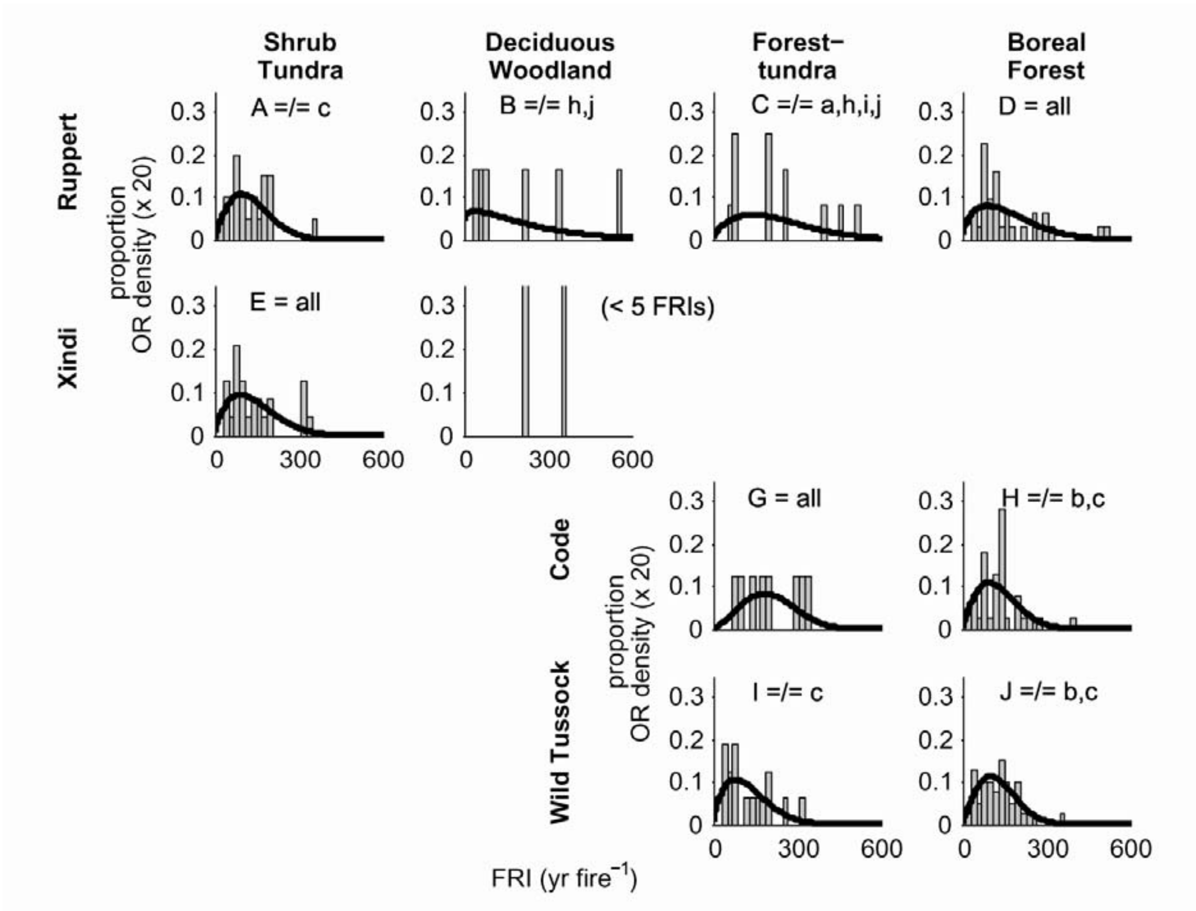
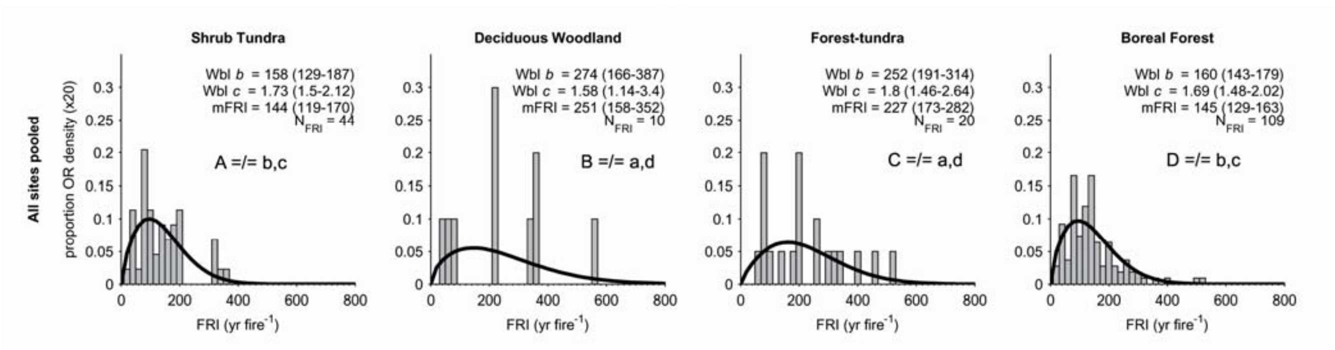


Figure 5.



Shrub Tundra
Deciduous Woodland
Forest−tundra
Boreal Forest
Ruppert
Xindi
Code
Wild Tussock
proportion OR density (x 20)
A =/= c
B =/= h,j
C =/= a,h,i,j
D = all
E = all
(< 5 FRIs)
G = all
H =/= b,c
I =/= c
J =/= b,c
0.3
0.2
0.1
0
0
300
600
FRI (yr fire^{-1})

Figure 6.


Shrub Tundra
Wbl b = 158 (129-187)
Wbl c = 1.73 (1.5-2.12)
mFRI = 144 (119-170)
N_FRI = 44
A =/= b,c
Deciduous Woodland
Wbl b = 274 (166-387)
Wbl c = 1.58 (1.14-3.4)
mFRI = 251 (158-352)
N_FRI = 10
B =/= a,d
Forest-tundra
Wbl b = 252 (191-314)
Wbl c = 1.8 (1.46-2.64)
mFRI = 227 (173-282)
N_FRI = 20
C =/= a,d
Boreal Forest
Wbl b = 160 (143-179)
Wbl c = 1.69 (1.48-2.02)
mFRI = 145 (129-163)
N_FRI = 109
D =/= b,c
All sites pooled
proportion OR density (x20)
FRI (yr fire-1)