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Variability of tundra fire regimes in Arctic Alaska: millennial-scale patterns and ecological implications

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Abstract. Tundra fires have important ecological impacts on vegetation, wildlife, permafrost, and carbon cycling, but the pattern and controls of historic tundra fire regimes are poorly understood. We use sediment records from four lakes to develop a 2000-yr fire and vegetation history in a highly flammable tundra region and compare this history with previously published fire records to examine spatial and temporal variability of tundra burning across Arctic Alaska. The four sites span a modern climatic gradient in the Noatak National Preserve, from warmer, drier down-valley locations to cooler, generally moister up-valley locations. Modern vegetation varies from herb- to shrub-dominated tundra from down- to up-valley sites, and pollen data suggest that this spatial pattern in vegetation persisted over the past two millennia. Peaks in macroscopic charcoal accumulation provide estimates of fire-event return intervals (FRIs), which did not vary significantly at millennial time scales but did vary across space. Down-valley sites burned relatively frequently over the past two millennia, with median FRIs of 150 years (95% CI 101–150) and FRI distributions statistically similar to those from ancient shrub tundra and modern boreal forest. At up-valley sites FRIs were significantly longer than those at down-valley sites, with a median FRI of 218 years (95% CI 128–285). These differences likely reflect the cooler growing-season temperatures and lower evaporative demand at up-valley sites, but local-scale variability in vegetation may have also shaped tundra fire regimes. Comparisons with other long-term fire records in Alaska reveal that the tundra biome can sustain a wide range of burning, with individual FRIs from as low as 30 years to more than 5000 years. These records together indicate that frequent tundra burning has occurred under a range of climatic and vegetation scenarios. The variety of tundra fire histories within Alaska suggests that the ecological impacts of tundra burning likewise vary widely, with important implications for wildlife-habitat maintenance and for the responses of tundra biophysical and biogeochemical processes to climatic change.

Key words: Alaska; Arctic; charcoal analysis; climatic change; environmental change; fire history; fire management; historical range of variability; Noatak National Preserve; pollen analysis; shrub tundra; tundra.

Introduction

Tundra fire regimes are a poorly understood component of a rapidly changing biome. Tundra burning impacts vegetation composition (Racine et al. 1987, 2004), nutrient cycling (Wookey et al. 2009), and permafrost dynamics (Shur and Jorgenson 2007), and it is an increasingly recognized feedback mechanism that links CO_2-induced warming to Arctic environmental change (e.g., Post et al. 2009, Mack et al. 2011). Tundra fires also represent an important management concern, for many of these same reasons and because they impact subsistence resources, including caribou populations (Jandt et al. 2008, Joly et al. 2010*a*). Despite the importance of tundra burning and the regional abundance of tundra fuels (e.g., ~33% of the state of Alaska), empirical information required for fire and resource management is lacking. For example, national fire initiatives such as the United State's LANDFIRE and Fire Regime Condition Class programs (*available online*)[6,7] require knowledge of historic fire return intervals, yet even this basic metric is missing in virtually all tundra ecosystems.

Increasing evidence suggests that Arctic warming may be affecting tundra fire regimes. In 2010, for example, 37 fires burned >43 180 ha in the Noatak National Preserve in northwestern Alaska (Fig. 1), the largest number of fires occurring in this area during a single year since record keeping began in 1950. Three years prior, a single event (the Anaktuvuk River Fire) on Alaska's North Slope more than doubled the total area burned north of 68° N in Alaska since Common Era

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[6] ⟨http://www.landfire.gov⟩
[7] ⟨http://www.frcc.gov⟩

(CE) 1950 (Fig. 1; Jones et al. 2009). This fire was associated with record-high temperatures, and record-low precipitation and sea-ice extent, and it marked the first time that the area had burned in at least 5000 years (Hu et al. 2010). Changes in fuel characteristics associated with increased shrub growth and/or density due to warming over the past several decades (e.g., Tape et al. 2006, Bhatt et al. 2010) may have contributed to these recent burns. Paleoecological evidence from "ancient" shrub tundra in the south-central Brooks Range (Fig. 1) indicates that highly flammable *Betula* (birch) shrubs that dominated ca. 13 000–11 000 calendar years before present (cal yr BP; Higuera et al. 2008) contributed to high levels of tundra burning during this period. In combination with studies documenting increased birch abundance in recently burned areas (e.g., Joly et al. 2010*a*), paleoecological patterns raise the possibility of positive feedbacks between increased shrub density and increased tundra burning.

Key vegetation and climatic controls thus appear to be changing in the tundra, which compounds the lack of understanding of tundra fire regimes. As in other ecosystems, the controls of tundra fire regimes can be largely understood through a conceptual framework that considers the limits of fire occurrence across multiple scales (Flannigan et al. 2009, Krawchuk et al. 2009, Parisien and Moritz 2009, Krawchuk and Moritz 2011). At landscape spatial scales and centennial-to-millennial time scales, low abundance of burnable biomass likely limits fire in the most environmentally extreme tundra ecosystem (e.g., barrens and high arctic tundra). Paleoecological evidence of frequent burning in ancient shrub tundra suggests that these fuel limitations can be removed if flammable biomass increases (Higuera et al. 2008). When or where sufficient biomass becomes available for burning, warm, dry, windy weather conditions promote vegetation flammability, fire ignition, and fire spread. In tundra ecosystems of Alaska, for example, over 95% of the variability in annual area burned over the past 60 years can be explained by interannual variations in growing-season temperature and precipitation alone (Hu et al. 2010). This implies that flammable tundra regions in the state have sufficient fuel to burn, but are limited more by fire-conducive climate.

The combined limits imposed by climate and vegetation interact (e.g., Littell et al. 2009) and result in spatial variability in tundra burning across Alaska. Whereas <1% of the North Slope has burned from CE 1950–2010, with 68% of that occurring in the Anaktuvuk River Fire of 2007 (Jones et al. 2009), ~8% and 13% of the Noatak National Preserve and Seward Peninsula has burned over this same period, respectively (Alaska Interagency Coordination Center [AICC], 1950–2009 fire perimeter data; ACIA 2004; Fig. 1). The paleoecological record also shows important contrasts among different tundra ecosystems, with frequent fires in ancient shrub tundra of interior Alaska (Higuera et al. 2008) contrasted with a lack of fire over the past 5000 years in areas on Alaska's North Slope (Hu et al. 2010). Differences in tundra burning on the modern landscape and in the past imply important differences in the limits of fire occurrence. As climate and vegetation change, it becomes increasingly important to reconcile these differences and understand how and why tundra fire regimes vary over space and time.

In this study we report on pollen records and the first long-term charcoal records from four lakes in the Noatak River watershed, one of the most flammable tundra ecosystems in Alaska (Fig. 1). These results are used to infer tundra fire history over the past 2000 years in the Noatak Valley. Focusing on this period represents a compromise between obtaining a paleo-data set sufficient for a statistical assessment of fire regimes (Baker 2009) and examining a period most relevant to the current vegetation and climate in the study region. To explore the general causes of spatial and temporal variability in tundra fire regimes, we compare our new fire reconstructions with published fire-history records from ancient and modern tundra in other regions of Alaska. Our results provide land managers with critical information on historic fire regimes and have important implications for understanding the controls of tundra fire regimes and anticipating impacts of ongoing environmental change in the Arctic.

Sample Lakes and Study Area

We reconstructed vegetation and fire history from sediments of four lakes spanning a 50-km east–west transect in the Noatak National Preserve, western Brooks Range, Alaska, USA (Table 1, Fig. 1). Climate in nearby Kotzebue (Fig. 1) is characterized by average maximum January and July temperatures of −15.51°C (SD = 4.41) and 15.09°C (SD = 1.71), respectively; mean annual precipitation is 245 mm (SD = 66), with 52% falling between July and September (Western Regional Climate Center, 1950–2009 observations; *available online*).[8] Regional vegetation is dominated by tussock–shrub tundra (including Cyperaceae [sedge], Poaceae [grass], *Betula* [birch], *Alnus* [alder], and *Salix* [willow] species); birch–Ericaceous shrub tundra; and willow shrub tundra, with white spruce (*Picea glauca*) reaching its up-valley limit along the Noatak River and adjacent drainages. Regional fire history since CE 1950 (Fig. 1) suggests fire cycles ranging from 175 to 193 years (Kobuk Ridges and Valleys Ecoregion; Kasischke et al. 2002, Joly et al. 2010*b*) to 480 years ("Noatak Lowlands"; Gabriel and Tande 1983).

Raven, Uchugrak, Poktovik, and Little Isac lakes (unofficial names) are small (3–6 ha), relatively deep (5–13 m) lakes with small, intermittent inlet and outlet streams (Table 1; Appendix A: Figs. A1 and A2). The two down-valley lakes (Raven and Uchugrak) are within

[8] ⟨http://www.wrcc.dri.edu/cgi-bin/cliMAIN.pl?ak5076⟩

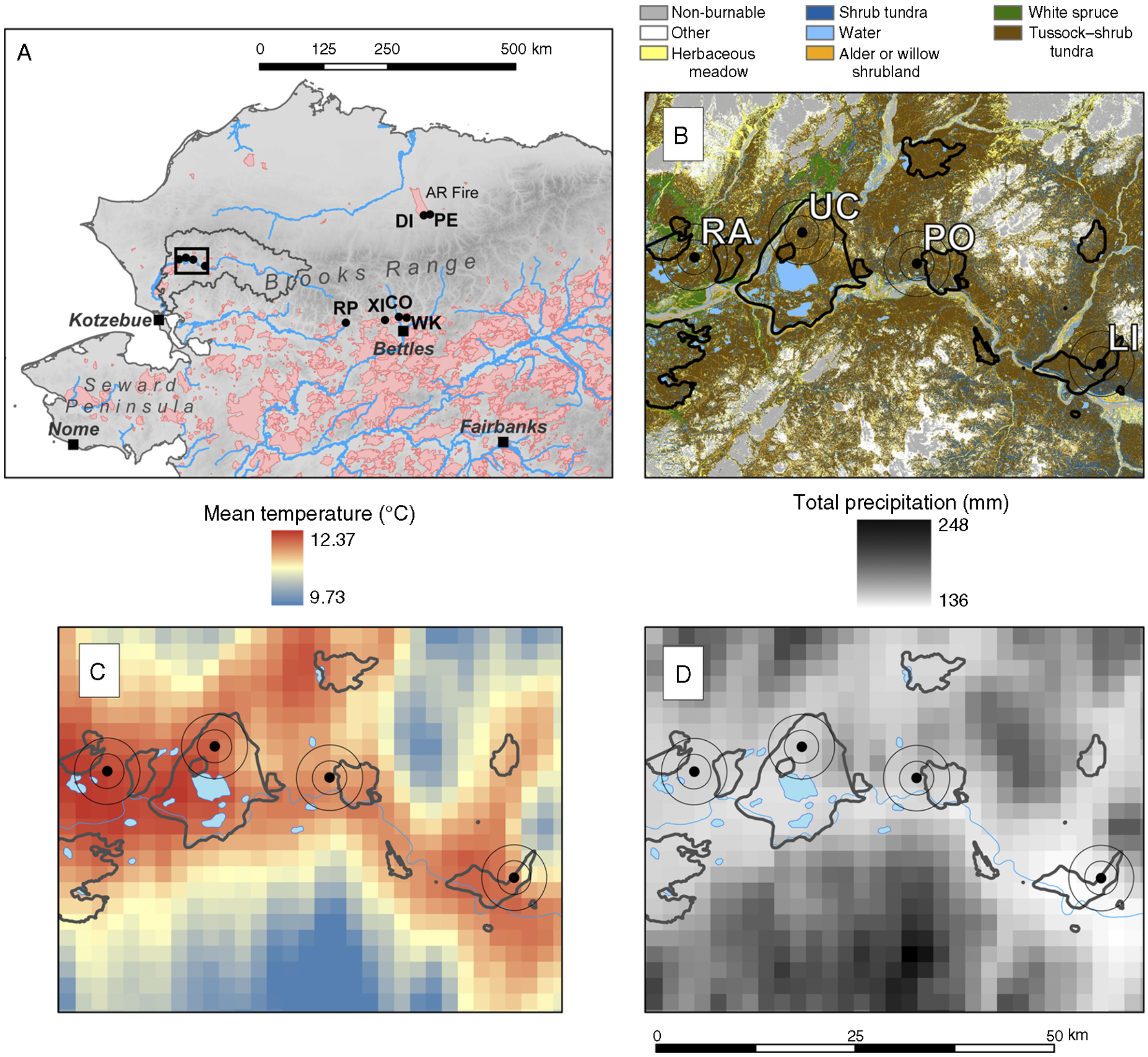

FIG. 1. Regional location, local vegetation, and local growing season climate (June–August) for the study area in the Noatak National Preserve, western Brooks Range, Alaska, USA. (A) Northern Alaska is shown with the black rectangle identifying the study area within the Noatak National Preserve (perimeter outlined in gray). The area burned from 1950 to 2009 is outlined in red polygons (Alaska Interagency Coordination Center; see footnote 9). Regional fire-history records referred to in the *Discussion* and Fig. 5 are also identified: Dimple (DI) and Perch (PE) lakes within the Anaktuvuk River Fire (AR) on the North Slope of the Brooks Range (Hu et al. 2010), and Ruppert (RP), Xindi (XI), Code (CO), and Wild Tussock (WK) lakes in the south-central Brooks Range (Higuera et al. 2009). (B) Vegetation in the Noatak study area is shown, summarized from Jorgenson et al. (2009) at 30-m resolution. The panel includes lakes used in this study: Raven (RA), Uchugrak (UC), Poktovik (PO), and Little Isac (LI). Fire perimeters corresponding to the four dots in the study area rectangle in panel (A) are shown as gray polygons. Circles around each lake have a 2 and 4 km radius, generous estimates for the spatial scale of charcoal and pollen records, respectively. (C) Mean June–August temperature and (D) total June–August precipitation at 2-km spatial resolution are shown, as estimated by the Parameter-elevation Regression on Independent Slopes Model (PRISM). Panels (C) and (D) also include fire perimeters and lake buffers as in panel (B).

the Kobuk Ridges and Valleys Ecoregion, whereas the two up-valley lakes are on the edge (Poktovik) or within (Little Isac) the Brooks Range Ecoregion (Nowacki et al. 2000). This division between the down-valley and up-valley regions is reflected by differences in local growing-season climate (June–August averages) and vegetation (within 4 km of each lake). As estimated by the Parameter-elevation Regression on Independent Slopes Model (PRISM; Daly et al. 1994), average temperature decreases by 0.4°C from western, down-valley sites to eastern, up-valley sites. Total precipitation is less variable among sites, with Little Isac Lake receiving less precipitation than other sites (Table 1, Fig. 1; Appendix A: Fig. A3). Temperature-based estimates of potential evapotranspiration suggest significantly higher evaporative demand at the two down-valley sites relative

TABLE 1. Lake characteristics and record quality from 2.5 ka BP to present in the Noatak National Preserve, western Brooks Range, Alaska, USA.

Characteristic	Site			
	Raven	Uchugrak	Poktovik	Little Isac
Site†				
Latitude	68°00′32.5″ N	68°03′07.4″ N	68°01′54.4″ N	67°56′27.9″ N
Longitude	162°02′08.5″ W	161°43′34.6″ W	161°22′29.0″ W	160°47′49.6″ W
Elevation (m above sea level)	118	216	160	210
JJA temperature (°C)	12.27 ± 0.07	12.10 ± 0.15	11.93 ± 0.10	11.87 ± 0.32
Total JJA precipitation (mm)	149 ± 1	153 ± 2	152 ± 2	146 ± 4
JJA PET‡ (mm)	155 ± 0.3	155 ± 1	154 ± 0.4	153 ± 1
JJA AET‡ (mm)	56 ± 1	57 ± 2	58 ± 3	55 ± 6
Surface area (ha)	5.7	3.7	6.3	2.9
Maximum water depth (m)	4.9	9.2	13.4	5.4
Record§				
Sedimentation rate (cm/yr)	0.0490 ± 0.3591	0.0476 ± 0.0960	0.0475 ± 0.1912	0.0206 ± 0.0367
Sample resolution (yr/sample)	11 ± 3	8 ± 1	21 ± 7	17 ± 6
Median sample resolution (yr/sample)	11	8	21	15
Median SNI¶	5.56	9.19	6.49	7.34

Notes: Note that sediment characteristics from 2.0 to 2.5 ka BP affect the determination of charcoal peaks within the period of analysis (2.0 ka BP to present). Variation around the mean is given as ±SD.

† June–August (JJA) climatology is from PRISM-derived data spanning 1961–1990, summarized across 25 2-km^2 pixels centered on each lake (representing an approximate radius of 4 km from each lake).

‡ Potential and actual evapotranspiration (PET and AET), estimated following Thornthwaite, as described in Willmott et al. (1985). AET estimates are based on a uniform available water capacity of 50 mm. See Appendix A for graphical representation of climate data.

§ Values are for the past 2500 yr.

¶ Signal-to-noise index (Kelly et al. 2010).

to up-valley sites, and estimates of actual evapotranspiration (AET) suggest lower AET at Little Isac Lake relative to all other lakes (Table 1; Appendix A: Fig. A3). Vegetation within 2–8 km of each lake also differs slightly among sites, with a higher abundance of tussock–shrub tundra at down-valley sites and higher abundance of low shrub tundra at up-valley sites; white spruce is present only at Raven Lake (Jorgenson et al. 2009; Fig. 1; Appendix A: Fig. A4).

Historic fire records from CE 1950–2007 (the year of sampling) include a 1300-ha fire that burned around Uchugrak Lake in CE 1977, and a 260-ha fire that burned around Little Isac Lake in CE 1984. Fires burned within 1.7 km of Raven Lake in CE 1972 and 1984, and within 0.7 km of Poktovik Lake in CE 1972 (Fig. 1; Alaska Interagency Coordination Center; data *available online*).[9]

METHODS

Lake sediments

Two overlapping sediment cores (5 or 8 cm diameter) were obtained from the deepest portion of each lake basin in the summer of 2007 (Appendix A: Fig. A2) using a modified Livingstone piston corer (Wright et al. 1984). An intact sediment–water interface was obtained from each lake using a polycarbonate tube fitted with a piston. The top 10–20 cm of the surface sediments were sliced at 0.5 cm in the field, and the remaining sections were split lengthwise in the laboratory and visually correlated based on distinct laminations. Correlated sections were sliced at 0.25 cm, including 3–5 cm of overlap between cores to confirm visual correlation with magnetic (not presented) and charcoal data. Sediments were dominated by dark black (Raven) or brown (all other lakes) gyttja, with intermittent clay-dominated laminations (<0.1–1 cm) or bands (1–3 cm). Basal sediments of each core were characterized by unlaminated clays, often with terrestrial debris (e.g., wood) intermixed.

Chronologies are based on the ^{210}Pb activity in the uppermost sediments and accelerator mass spectrometry (AMS) ^{14}C ages of terrestrial macrofossils from deeper sediments. Sediments for ^{210}Pb analysis were processed following Eakins and Morrison (1978), and activity was measured with an Octête Plus alpha spectrometer (Ortec, Oak Ridge, Tennessee, USA) at the University of Illinois (Champaign, Illinois, USA). Sample age was estimated using a constant-rate-of-supply (CRS) model with an old-age correction adapted from Binford (1990). Terrestrial macrofossils were treated with an acid–base–acid procedure (Oswald et al. 2005) and submitted to Lawrence Livermore National Laboratory (Livermore, California, USA) for radiocarbon measurements. All ^{14}C ages were calibrated to years before present (BP, before CE 1950) using the IntCal 04 data set in CALIB version 5.01 (Reimer et al. 2004). Age models were developed using a weighted cubic smoothing spline function in MATLAB with confidence intervals estimated through Monte Carlo methods (following Higuera et al. 2009).

[9] ⟨http://fire.ak.blm.gov/predsvcs/maps.php⟩

For charcoal analysis, subsamples of 2–10 cm^3 (mode = 5 cm^3) were taken from continuous core slices, freeze-dried to remove moisture, and placed in a 16 mL solution of 5.25% bleach and 10% sodium metaphosphate to soak overnight (<24 h). The treated sediment was gently washed with distilled water through a 180-μm sieve, and charcoal was identified at 10–40× magnification. Charcoal concentrations (pieces/cm^3) were multiplied by the sediment accumulation rate (cm/yr) to calculate charcoal accumulation rates (CHAR, pieces·cm^{-2}·yr^{-1}).

Pollen identification was performed on 1-cm^3 subsamples at varying intervals. Sample preparation followed standard protocols for lake sediments (Faegri et al. 1989), and pollen grains were counted at 40–100× magnification. Pollen data were expressed as a percentage of total terrestrial pollen grains.

Statistical treatment of pollen and charcoal data

Pollen percentages from the seven most dominant taxa (*Picea*, *Betula*, *Alnus*, Cyperaceae, Poaceae, Ericaceae, and Chenopodiaceae) were compared for the past 2000 years, visually in boxplots and continuously in time-series plots fit to individual pollen samples. Prior to statistical analyses, charcoal samples were interpolated to 15 years, approximately equal to the median sample resolution of each record (Table 1). To infer the timing of local fires, we decomposed charcoal records to identify distinct peaks by applying a uniform set of threshold criteria to interpolated CHAR series, C_{int}, using the CharAnalysis program (version 1.1, *available online*; as in Higuera et al. 2009).[10] We use "local" to refer to distances within 500–1000 m of each lake, representing an area of 100–300 ha (1–3 km^2; Gavin et al. 2003, Lynch et al. 2004, Higuera et al. 2007). Low-frequency trends in CHAR ("background," C_{back}) were estimated using a 750-yr, locally weighted regression robust to outliers (Cleveland 1979). C_{back} was removed from the interpolated series by subtraction, creating a detrended "peak" series, C_{peak} (i.e., $C_{peak} = C_{int} - C_{back}$). We used a Gaussian mixture model to separate C_{peak} into two components: C_{noise}, variations around background that reflect natural and analytical effects (e.g., distant fires, sampling variability), and C_{fire}, variations exceeding variability in C_{noise}, assumed to reflect one or more local fires within the 15-yr interval (hereafter "fire event"). Specifically, the 99th percentile of the estimated C_{noise} distribution served as the threshold value separating C_{fire} from C_{noise}. As in Higuera et al. (2009), all peaks were screened to test whether variations between a peak and the smallest non-peak sample within the previous 10 samples (i.e., 150 years) differed statistically based on the charcoal counts and sample volume. Samples were considered different if the minimum-count test yielded a P value <0.25. This liberal cutoff was necessary given low charcoal concentrations in tundra lakes (see *Discussion*).

Quantifying and comparing fire regimes

We inferred 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 given peak (i.e., all samples above the threshold value; pieces·cm^{-2}·$peak^{-1}$), is a measure of total charcoal deposition per fire event (e.g., Whitlock et al. 2006, Higuera et al. 2009, Hély et al. 2010). Systematic differences in peak magnitude between sites were used as a qualitative proxy for average biomass consumed per fire, which should reflect fire size and/or fuel consumption for a given area burned. Interpretations rest on theoretical relationships between fire and charcoal deposition at a lake (Higuera et al. 2007), although links between fuel consumption and peak magnitude have yet to be tested empirically. To test the null hypothesis that peak magnitude values did not differ among sites, we used a nonparametric analysis of variance (Kruskal-Wallis test) with an alpha of 0.05. When the null hypothesis was rejected, we applied a multiple comparison test to isolate the source of significance.

We reconstructed the frequency component of past fire regimes with fire-event return intervals (FRI, years), calculated from the time elapsed between any two consecutive charcoal peaks. To facilitate statistical comparisons among sites and with previous studies, we modeled the distribution of FRIs from the past 2000 years with a two-parameter Weibull model, using maximum-likelihood techniques (e.g., Johnson and Gutsell 1994); this procedure included a Kolmogorov-Smirnov test to evaluate goodness of fit between the modeled parameters and the empirical FRIs (as described by Higuera et al. 2009). This approach assumes that variations in climate (e.g., Little Ice Age; Clegg and Hu 2010) and vegetation (see *Results*) affecting the probability of fire occurrence over the past 2000 years were not large enough to confound our goal of obtaining an adequate sample size to quantify fire frequency that is relevant to current fire management (evaluated in the *Results*). The null hypothesis that FRI distributions from any two sites did not differ was tested using a likelihood-ratio test based on the Weibull b and c parameters, with an alpha of 0.10 (Higuera et al. 2009). If FRI distributions at adjacent sites did not differ significantly, then FRIs were pooled to form a composite distribution and comparisons were repeated. In addition to this discrete view of fire occurrence, we also calculated the mean FRI over continuous 2000-yr time intervals, and then smoothed these series with a 2000-yr locally weighted regression. This necessitated the use of fire-history information from before 2000 yr BP, which was derived using the same parameters described previously, and is

[10] ⟨http://code.google.com/p/charanalysis/⟩

based on extrapolated data for the past 1000 years. The intent of this step is to display long-term trends in mean FRIs. For both the discrete and continuous displays of fire occurrence, 95% confidence intervals were estimated from 1000 bootstrap samples from the (local or global) FRI distributions. All data and statistical analyses were done using MATLAB software (2009 a–b; Mathworks, Natick, Massachusetts, USA).

To place fire history into the context of other arctic and boreal fire regimes, we compared FRI distributions and peak magnitudes from this study to the same metrics from sites reflecting "ancient herb tundra," "ancient shrub tundra," and "modern boreal forest" in the south-central Brooks Range (Higuera et al. 2008, 2009), and "modern tundra" on the central North Slope (Hu et al. 2010; Fig. 1). Two of the comparison records (Ruppert and Xindi) represent fire regimes from ancient herb tundra, characterized by sparse vegetation dominated by Cyperaceae, *Salix*, Poaceae, and *Artemisia* that existed under climatic conditions that were much cooler and drier than present, ca. 15.5–13.3 kiloannum (ka) BP. The same two sites also provide information on fire regimes from ancient shrub tundra, a more biomass-rich vegetation dominated by shrub *Betula*, and notably with no *Alnus*, during climatic conditions that were likely cooler and drier than present, ca. 13.3–10.3 ka BP. Three records (Ruppert, Code, and Wild Tussock) reflect modern boreal forest, characterized by *Picea mariana*-dominated boreal forest from 5.5 ka BP to present. The final two records come from the only other study to reconstruct fire history in modern tundra environments (Hu et al. 2010), which detect a single fire (in CE 2007) over the past 1.2 and 5.0 ka. Tundra in this region is generally less shrub-dominated than tundra in the Noatak study area, but it includes many of the same species (Walker et al. 2005).

Results

Chronologies

A total of 35 ^{14}C dates and 80 ^{210}Pb samples were obtained to develop age–depth models for the four records, with 1–3 ^{14}C dates constraining the last 2.5 ka at each lake (Fig. 2; Appendix B: Table B1 and Fig. B1). We present the age–depth models back to 3.2 ka BP (Fig. 2) because the cubic spline near 2.5 ka BP is affected by samples older than 2.5 ka BP. Overall, well-developed ^{210}Pb chronologies allowed us to compare recent peaks with known fires, but the precise timing of older peaks should be interpreted with caution because of ^{14}C chronological uncertainties. At Little Isac Lake, three ^{14}C dates, at 22.25, 49.25, and 74.75 cm, were not used in the chronology because they were anomalously old (top two) or young (bottom-most) and inconsistent with the nine other ^{14}C dates in the core. At Poktovik Lake, two ^{14}C dates were not used because they were anomalously old (at 32.50 cm) or young (at 97.25 cm) and also inconsistent with nine other ^{14}C dates in the core (Fig. 2; Appendix B: Table B1 and Fig. B1). The anomalously old ages likely resulted from long terrestrial residence times of the dated macrofossils before deposition in lake sediments (Oswald et al. 2005). We do not have a clear explanation for the anomalously young ages, although one sample was near core-drive end (3.5 cm) and may have contained young organic material. The paucity of terrestrial plant macrofossils in our cores made it infeasible to obtain additional ^{14}C ages. Such chronological issues are common to arctic and subarctic sediment records (e.g., Oswald et al. 2005, Hu et al. 2010), and given our focus on comparisons over a 2000-yr period, our results remain robust to small chronological variations.

Sediment accumulation rates since 2.5 ka BP averaged 0.0308 cm/yr (standard deviation = 0.0301; median = 0.0248, range 0.0081–0.5652). Sample resolutions for charcoal analysis averaged 12 yr/sample (median = 11, range 1–32), with 87% of the samples representing sediment deposition of $\leq$15 years (i.e., the interpolation interval for charcoal analysis; Table 1, Fig. 2). A 15-yr resolution implies a minimum detectable mean fire return interval of 30 years and mean fire return interval of ~75–125 years (i.e., ~3.3–5.0 times sample interval; Clark 1988, Higuera et al. 2007). Given repeated burning within 30-yr intervals in the study area and elsewhere in Alaska (Fig. 1), we use the term "fire event" to acknowledge that some peaks may include more than one fire.

Pollen records

A total of 31 pollen samples were counted across the four sites from 2.5 ka BP to present, with an average of 347 ± 195 (SD) yr between samples (range 59–797) and 285 ± 11 terrestrial grains/sample (range 258–306). Pollen spectra included over 28 terrestrial taxa but were dominated by *Alnus*, *Betula*, Cyperaceae, and Poaceae, which together accounted for >63% of terrestrial pollen grains in all but one sample from all records. With *Picea*, Ericaceae, and Chenopodiaceae, these seven taxa made up 84–95% of the terrestrial pollen grains in all but one sample (from Little Isac Lake, where they made up 66% of terrestrial pollen grains). Other taxa included *Populus*, *Salix*, and *Artemisia*.

Pollen spectra changed little since 2.5 ka BP (Appendix B: Fig. B2), and among-site differences were broadly consistent with the greater modern abundance of *Picea glauca* and tussock–shrub tundra near down-valley sites, and the greater abundance of shrub tundra near up-valley sites (Appendix A: Fig. A4; Fig. 3). The relative proportion of *Picea* pollen (assumed to be *P. glauca*) was highest at Raven Lake (median = 8%) in the lower Noatak Valley, more than twice that at other sites (medians from 3–4%). Relative to other sites, Little Isac Lake had lower proportions of *Alnus* (median = 22% vs. 29–32%) and Cyperaceae (17% vs. 23–26%), and higher percentages of *Betula* (median = 23% vs. 12–17%), Ericaceae (median = 5% vs. 2–4%), and Chenopodiaceae (median = 5% vs. 2–3%).

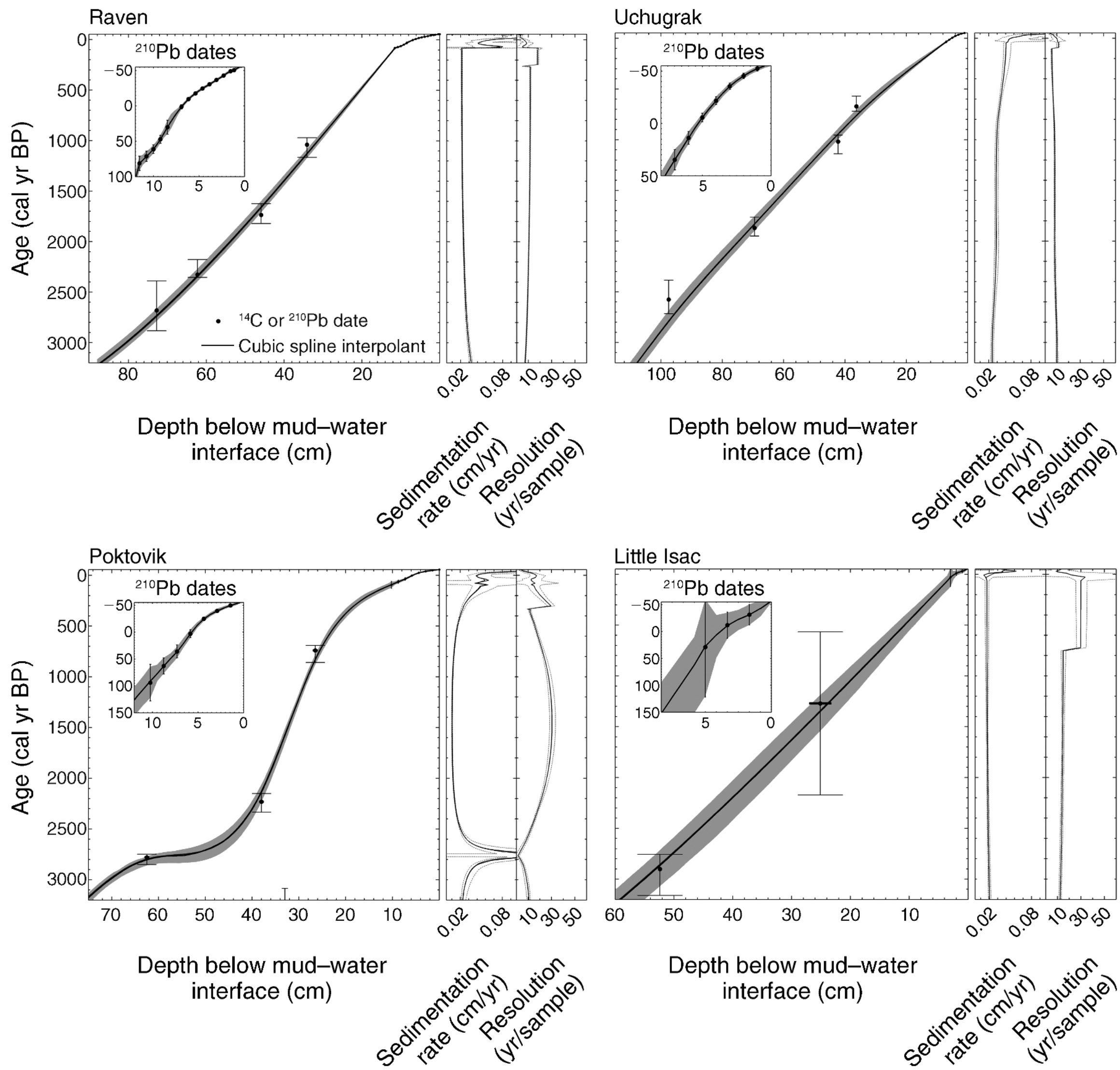

FIG. 2. Age–depth models since 3.2 ka BP with radiometric ages (^{210}Pb and ^{14}C) and the cubic spline fit, sedimentation rate (Sed.), and sample resolution, all with 95% confidence intervals. Axes of the inset panels match those of the main panels but are zoomed in at the top of the *y*-axis scale to show the ^{210}Pb dates. Confidence intervals were based on Monte Carlo resampling of 1000 chronologies, with the age of each sample randomly selected from the probability distribution of each calibrated ^{14}C and ^{210}Pb date (Higuera et al. 2009). Sample resolution is generally <30 years per sample, providing suitable records for high-resolution charcoal analysis. Record ages vary from 6000 to ca. 12 000 years, and the full age–depth models are shown in Appendix B: Fig. B1. Note that the period of charcoal analysis in this study starts at 2500 cal yr BP; thus the rapid sedimentation rate at Poktovik Lake, ca. 2800 cal yr BP, does not influence the record (Fig. 4). Error bars and shaded areas show 95% CI.

Charcoal records and inferred fire history

Charcoal abundance was low, with concentrations varying between 0 and 27 pieces/cm^3 (corresponding to counts between 0 and 80 pieces; Appendix B: Fig. B3). The mean (median) charcoal count in each sample at Raven, Uchugrak, Poktovik, and Little Isac lakes was 2 (1), 1 (0), 2 (0), and 9 (4) pieces, respectively. The corresponding mean (median) values of charcoal accumulation rates (CHAR) are 0.0165 (0.007), 0.013 (0.000), 0.009 (0.002), and 0.053 (0.027) pieces·cm^{-2}·yr^{-1} (Figs. 4 and 5; Appendix B: Fig. B3).

Despite low charcoal abundance, two aspects of our records support the interpretation of identified peaks as indicators of fire events within (and not beyond) 1 km of each lake. First, documented fires that burned around Uchugrak and Little Isac lakes in CE 1977 and 1984, respectively, were detected by peaks spanning CE 1977–1992 at both sites (Fig. 4). Conversely, documented fires burning 1.0–1.7 km from Raven and Poktovik lakes

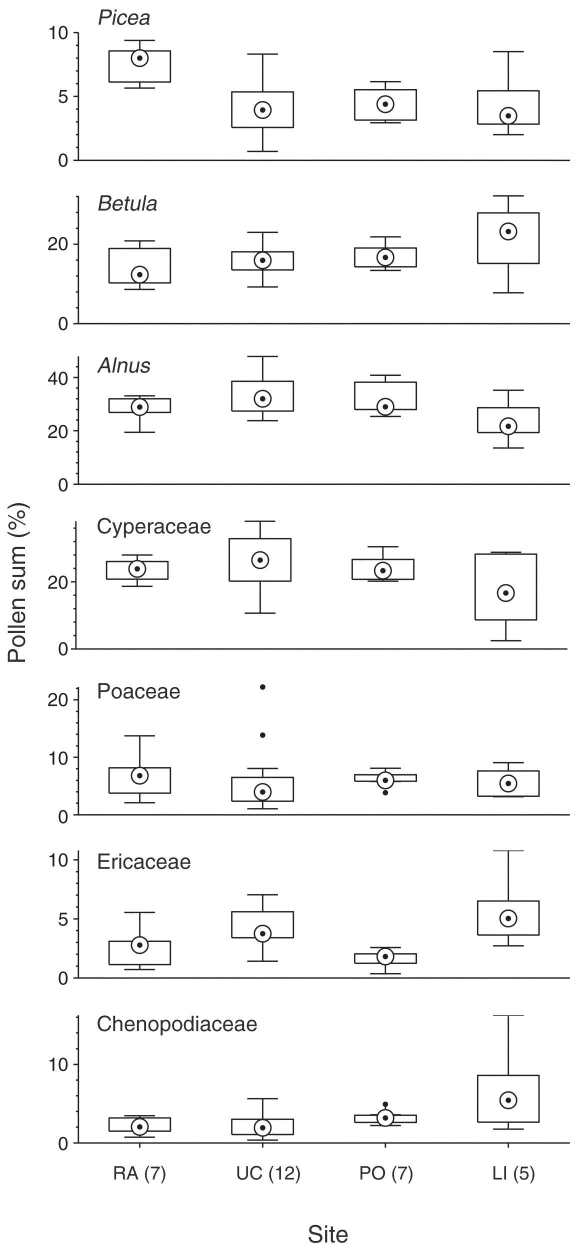

FIG. 3. Boxplots of pollen percentages from the seven most dominant taxa at each site, from 2.5 ka BP to present. Boxes outline the 25th and 75th quartiles, circled dots identify the median (50th quartile), whiskers capture ~90–95% of the values, and dots identify outliers (values beyond the 90–95th quartile). Numbers in parentheses on the *x*-axis are the number of pollen samples at each site; see Fig. 1 for lake abbreviations.

were *not* detected. While these nearby fires contributed charcoal to each record, neither peak passed the minimum-count screening, lending support to this criterion. Second, CHAR time series exhibited significant high-frequency variability, with median signal-to-noise index values (as defined and described by Kelly et al. 2011) greater than three, the theoretical minimum value for justifying peak analysis (Table 1).

A total of 13, 16, 9, and 7 fire events were identified at Raven, Uchugrak, Poktovik, and Little Isac lakes over the past 2000 years, respectively, with FRIs varying from 30 to 840 years (Table 2; additional fire events were detected prior to 2 ka BP). Despite this large variability, millennial-scale mean FRIs did not vary significantly through time at any site, as inferred from overlapping 95% confidence intervals (Fig. 4D). When summarized at the site level, FRI distributions did not differ between the two down-valley sites (Raven and Uchugrak; $P = 0.89$), or between the two up-valley sites (Poktovik and Little Isac; $P = 0.53$; Table 2). When pooled into two populations, however, FRIs from the two down-valley sites were significantly shorter than those from the two up-valley sites, with the former burning nearly twice as often as the latter on average ($P = 0.018$). Specifically, the pooled distribution from Raven and Uchugrak lakes had a mean and median (95% confidence interval) of 142 years (115–174) and 150 years (101–150), respectively, whereas the pooled distribution from Poktovik and Little Isac lakes had a mean and median of 263 years (175–374) and 218 years (128–285), respectively (Table 2).

Comparisons among regional FRIs indicate that the Raven-Uchugrak distribution did not differ from the FRI distributions of ancient shrub tundra ($P = 0.694$–0.941) or modern boreal forest in the south-central Brooks Range ($P = 0.291$–0.989). The Poktovik-Little Isac distribution, in contrast, had significantly longer FRIs as compared to ancient shrub tundra ($P = 0.065$–0.069) and two of the three sites from modern boreal forest ($P = 0.004$–0.011; Fig. 5).

Peak magnitudes varied across all sites from 0.006 (Raven) to 48.92 (Little Isac, maximum value of sample not plotted), with mean (median) values of 0.65 (0.26), 0.75 (0.40), 0.57 (0.04), and 8.32 (1.10) pieces·cm^{-2}·peak^{-1} at Raven, Uchugrak, Poktovik, and Little Isac lakes, respectively (Fig. 5B). A Kruskal-Wallis ANOVA suggests significant differences among sites ($P = 0.017$); multiple comparisons indicate that Little Isac had higher peak magnitudes than Poktovik, with no other between-site differences. Comparisons among regional records also indicate significant differences ($P < 0.001$). With the exception of Little Isac Lake, peak magnitudes in the study area were significantly lower than two to five sites from ancient shrub tundra and/or modern boreal forest (Fig. 5B).

DISCUSSION

Our study provides the first millennial-scale records of fire history in the region and contributes to a growing understanding of fire regimes in tundra ecosystems. Inferred fire-event return intervals (FRIs) indicate that tundra has sustained frequent burning in the past, and this history provides land managers with the information needed to make informed management decisions and comply with federally mandated programs. Variability among our study sites indicates the relative roles of climate and vegetation in shaping tundra fire regimes,

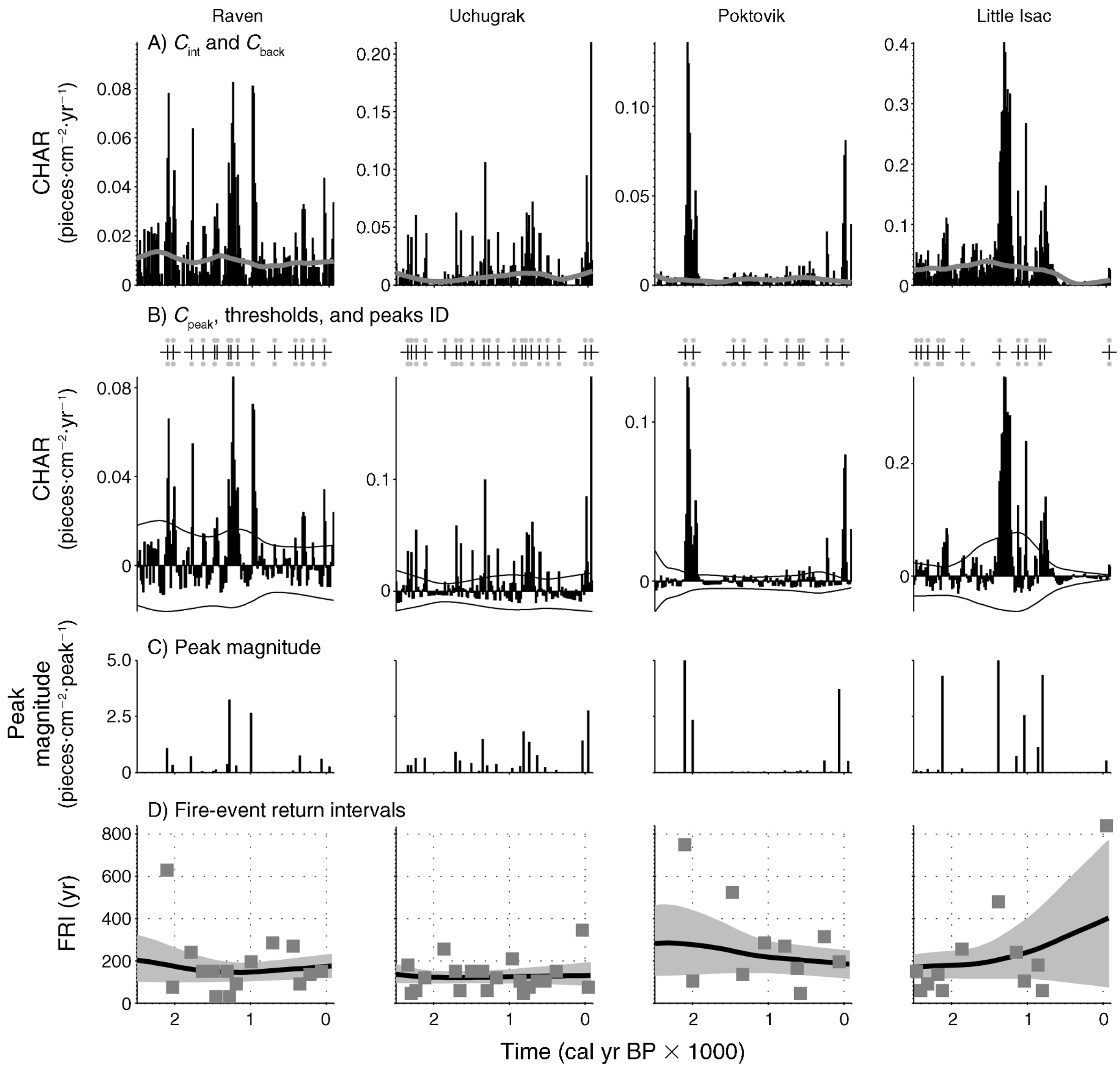

FIG. 4. Charcoal records, peak analysis, and inferred fire history. (A) Interpolated charcoal record (C_{int}) and 750-yr trends defining background charcoal (C_{back}); CHAR is the charcoal accumulation rate. (B) Peak charcoal (C_{peak}, defined as $C_{int} - C_{back}$), locally defined thresholds for peak detection, and peaks identified (ID) from samples exceeding the positive threshold and passing the minimum-count screening (see *Methods*). Peaks identified by three threshold levels are shown above the record (gray dots), with those used for final interpretation identified by the "+" symbol. (C) Peak magnitude, integrated charcoal accumulation from each sample exceeding the positive threshold in panel (B). (D) Fire-event return intervals, calculated as the time elapsed between consecutive peaks identified in panel (B). The dark black line is the 2000-yr mean fire-event return interval (FRI), and the gray envelope is the 90% confidence interval around the mean FRI, based on 1000 bootstrapped samples of FRIs, with replacement.

and comparisons to other systems have important implications for fire ecology in tundra ecosystems, land management, and environmental change in the Arctic.

Reconstructing tundra fire regimes with lake-sediment records

Lake-sediment charcoal records offer one of the only tools available to infer past tundra fires regimes, but this method comes with several limitations, and its use in tundra ecosystems is still in its infancy (e.g., Higuera et al. 2008, Hu et al. 2010). In combination with records reported by Hu et al. (2010), our records suggest that modern tundra fires produce enough charcoal to leave distinct evidence of past fire occurrence, despite lower biomass consumption than in forested ecosystems. Comparisons among the uppermost peaks in each record and fires from the observational record support the interpretation that charcoal peaks represent fires within and not beyond ~1.0 km of each lake (Figs. 1 and 4). Additionally, high-frequency variability in charcoal accumulation rates (Fig. 4) was statistically

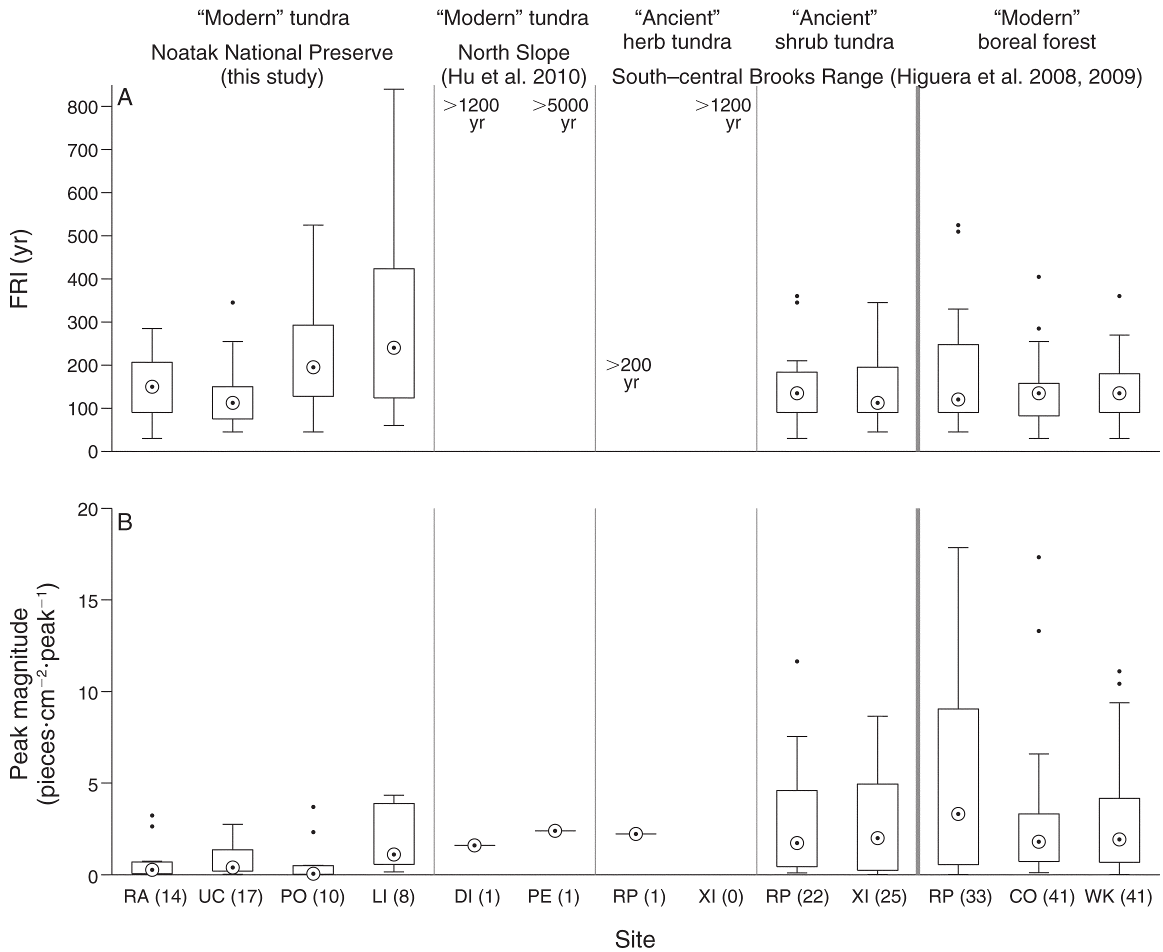

FIG. 5. Among-site and among-study comparisons of fire history metrics. (A) Boxplots (as in Fig. 3) of fire-event return intervals from (left to right) this study, records within the 2007 North Slope Anaktuvuk River Fire, "ancient" herb tundra from the late Pleistocene, "ancient" shrub tundra from the late Pleistocene and early Holocene, and "modern" boreal forest from the past 5000 years. Boxplots represent observed fire return intervals; numbers (e.g., >1200 yr) are given in the absence of observed FRI. (B) Charcoal peak magnitude from these same records. Numbers in parentheses on the x-axis are the number of fire events at each site; see Fig. 1 for lake abbreviations.

distinct from a null model of random variability (i.e., signal-to-noise index; Kelly et al. 2011; Table 1).

While these aspects provide confidence in our estimated FRIs, analyzing sediment records with low charcoal production is a particular challenge in tundra ecosystems. Even when increasing sediment sample sizes beyond volumes typically used in other ecosystems such as boreal or temperate forests (5–10 cm^3 vs. 1–3 cm^3), low

TABLE 2. Fire history statistics from 2.0 ka BP to present at each site.

Site(s)	n_{FRI}	Fire-history parameter (95% CI): Range of fire-event return intervals (yr)	Mean fire-event return interval (yr)	Median fire-event return interval (yr)	Weibull b parameter (yr)
Raven	13	30–285	151 (108–199)	150 (90–195)	170 (120–217
Uchugrak	16	45–345	135 (98–176)	113 (75–150)	153 (114–198)
Poktovik	9	45–525	227 (139–327)	195 (105–300)	255 (165–372)
Little Isac	7	60–840	309 (134–521)	240 (105–548)	337 (169–576)
Raven + Uchugrak	29	30–345	142 (115–174)	150 (101–150)	161 (130–197)
Poktovik + Little Isac	16	45–840	263 (175–374)	218 (128–285)	291 (198–410)

Notes: Confidence intervals were estimated through 1000 bootstrapped samples of the sample population, with replacement; n_{FRI} is the number of fire-event return intervals.

† The fire cycle, FC, is equivalent to the point-specific mean fire return interval (Johnson 1994), and it can be calculated from the Weibull model as: $FC = b\Gamma(1/c + 1)$, where Γ is a gamma function. To the extent that the Weibull model fits the observed data well, then FC will be equivalent to the calculated mean fire (event) return interval.

charcoal counts were common in our samples (Appendix B: Fig. B3). This introduces additional variability into the peak detection process (Higuera et al. 2010), which in turn can lead to falsely inferred fires (i.e., false positives). We account for some of this variability and guard against false positives with our minimum-count screening, which removed small peaks that would otherwise be included in FRI statistics. Future fire history studies in tundra should work to maximize the number of charcoal pieces in each sample, possibly by sieving for smaller and thus more abundant size fractions (e.g., >125 or 150 μm) or increasing the volume of sediment subsampled.

Patterns and drivers of past tundra burning in the Noatak study area

Over the past two millennia, temporal variability in the long-term mean FRI was statistically insignificant, driven largely by one or two (long) FRIs (Fig. 4). This lack of change through time suggests an insensitivity of tundra fire regimes to regional climatic variability (e.g., Little Ice Age [LIA] cooling, ca. CE 1500–1800; Barclay et al. 2009, Clegg and Hu 2010). However, this interpretation is inconsistent with variable burning in relation to climatic conditions across our sites (discussed in the next paragraph) and with previous studies inferring fire-regime shifts in southern Alaskan boreal forest (Tinner et al. 2008) in response to temperature variations of 1–2°C during the LIA (e.g., Hu et al. 2001, Clegg et al. 2010). Climatic variability in the Noatak may have been more subtle than in southern Alaska, but the lack of change in our records also reflects low statistical power for detecting all but the most extreme FRI shifts from centennial-scale climatic variability. For example, even with four records spanning the 300-yr LIA, statistical power for detecting a 30% decrease or 50% increase in mean FRI (e.g., from ~150 to 100 yr, or from 150 to 225 yr) is only ~0.30 (Higuera 2006).

In contrast to the lack of significant change over time, spatial variability was evident in the long-term FRIs. The down-valley and up-valley sites, which are currently in different ecoregions, burned at significantly different rates (Table 2, Figs. 4 and 5), suggesting local-scale variability as a driver of tundra fire regimes over the past two millennia. Shorter mean FRIs at down-valley (142 years, 95% CI = 115–174) compared to up-valley (263 years, 95% CI = 175–374; Table 2) sites coincide with a higher growing-season (June–August) temperature (0.4°C) at the down-valley sites today (Table 1, Fig. 1; Appendix A: Fig. A3). Precipitation is less variable among sites, with the exception of Little Isac Lake, which has lower growing-season precipitation than the other sites (Table 1, Fig. 1; Appendix A: Fig. A3). Furthermore, limited data on lightning strikes in our study area show no evidence suggesting that greater fire occurrence at down-valley locations is driven by greater lightning density (strikes per unit area per unit time). This current climatic gradient (mainly temperature) likely persisted over the past two millennia, directly enhancing fire risk at down-valley sites by increasing potential evapotranspiration (Table 1; Appendix A: Fig. A3), reducing fuel moisture, and thus facilitating fire ignition and spread. Although paleoclimate data needed to evaluate the role of spatial patterns of climate across our study area are lacking, our interpretations are consistent with the influence of seasonal and annual climate on tundra burning in the recent past. Large tundra fires (Wein 1976, Racine et al. 1985, Jones et al. 2009) and annual tundra area burned (Hu et al. 2010) occur under extreme warm, dry summer conditions, which increase fire risk through drying of fuels and facilitating fire spread. Furthermore, area burned in Alaskan tundra has been biased toward regions with warm, dry summer climate (Higuera et al. 2008), likely due to a combined effect of climate on vegetation and the frequency of fire weather. Climate–fire relationships at both seasonal to decadal time scales thus suggest that the climatic gradient in the Noatak study area is a plausible explanation for the spatial variability in past fire occurrence. Although this gradient is small, the persistent difference between sites seems to have been sufficient to create differing fire regimes.

In addition to their direct impacts, mean growing-season temperature and precipitation also strongly influence tundra vegetation (Walker et al. 2005), which in turn can affect fire risk by determining the nature and abundance of fuel available for burning. More frequent burning at down-valley compared to up-valley sites is consistent with the higher surface-area-to-volume ratio and reproductive traits of tussock tundra (e.g., resprouting or reseeding potential; Wein and Bliss 1973) and the flammable nature of the sparse but present coniferous fuels. Both traits increase the probability of fuel drying and fire spread when ignitions occur (e.g., Rothermel 1983). Actual evapotranspiration (AET), a water-balance metric that integrates the effects of temperature and precipitation, is broadly linked to primary productivity (see review by Fisher et al. 2011). Lower AET at Little Isac Lake is consistent with lower plant productivity, potentially limiting frequent burning due to low fuel abundance. Thus, the across-site variability in

Table 2. Extended.

Weibull *c* parameter (unitless)	Weibull-estimated fire cycle†
2.0 (1.38–3.71)	151 (111–192)
1.89 (1.53–3.21)	136 (100–178)
1.76 (1.28–4.12)	227 (147–330)
1.31 (1.00–3.12)	311 (147–504)
1.93 (1.56–2.58)	143 (114–172)
1.43 (1.15–2.48)	264 (179–370)

burning also suggests a possible link between vegetation (composition and abundance) and fire. Untangling the relative effects of climate vs. vegetation on tundra fire regimes is difficult because fire itself shapes tundra species composition (Racine et al. 2004, 2006, Lantz et al. 2010) and the impacts of fire on tundra vegetation can interact with climate (de Groot and Wein 1999). For example, warmer growing-season temperatures favor increased shrub growth or abundance (e.g., Chapin et al. 1995, Walker et al. 2005, 2006), yet both modern (Fig. 1) and paleo- (Fig. 3; Appendix B: Fig. B1) vegetation records indicate great shrub abundance at cooler, up-valley sites in our study area. This variability may reflect more frequent burning favoring tussock over shrub growth forms at down-valley sites (Wein and Bliss 1973, de Groot and Wein 2004).

The patterns of burning in our study area thus leave open several alternative hypotheses: (1) spatial climatic variability (primarily temperature) largely determined fire frequency, which in turn influenced vegetation composition, (2) spatial climatic variability (as influencing AET) largely determined tundra vegetation, which in turn influenced fire frequencies through patterns of fuel abundance and flammability, or, most likely, (3) some combination of (1) and (2) determined historic fire regimes. Evaluating these alternative hypotheses requires detailed analyses of tundra vegetation across gradients of climate and fire (in space or time), through either manipulative or natural experiments. Human activity in the Noatak Valley is an unlikely cause of the spatially variable fire regimes. For example, while human occupation in the region dates to at least 13 ka BP (Rasic 2011), populations are thought to have been small and thinly dispersed, and there is little evidence suggesting that prehistoric cultures used fire nondomestically (J. Rasic, *personal communication*). What is clear from our data is that tundra ecosystems in the Noatak have burned relatively frequently in the past, with rates that varied across space in ways consistent with spatial patterns in growing-season climate and local vegetation.

Variability in tundra fire regimes across space and time

Our results add to a growing body of evidence indicating that not only can tundra burn (e.g., Wein 1976, Racine et al. 1985, Jones et al. 2009), but it can do so at rates much higher than observed across most of the tundra biome (e.g., Wein 1976, Whelan 1995). Previously, the only long-term evidence of frequent tundra burning in Alaska was from "ancient" shrub tundra of the late Pleistocene and early Holocene (Tinner et al. 2006, Higuera et al. 2008), a biome that was likely shrubbier than modern tundra, lacked the presence of *Alnus,* and existed under climatic conditions that were likely cooler and drier than present. Given cooler-than-present climate, the expansion of shrub birch was likely a key factor facilitating frequent burning in the ancient shrub tundra by adding flammable fuel to the landscape (Higuera et al. 2008). Similar FRIs among the two down-valley sites reported here and ancient shrub tundra (Fig. 5) indicate that frequent burning can occur under a range of tundra vegetation scenarios. Abundant birch shrubs in the ancient shrub tundra were sufficient for frequent burning, but so too was the tussock–shrub tundra that dominated our down-valley sites. In both cases, climate and vegetation supported frequent burning, but their relative influence likely differed. Our records thus refute a simple relationship between fire occurrence and climate or species composition. Rather, they point to the interacting effects of climate and biomass (type and abundance) in facilitating frequent burning in tundra ecosystems and suggest that trade-offs between these two can lead to tundra burning under diverse climatic and vegetation scenarios.

Fire histories from the Noatak study area and ancient shrub tundra stand in stark contrast to those in modern tundra on Alaska's North Slope (Fig. 1; Hu et al. 2010) and ancient herb tundra (Higuera et al. 2009), which reveal little or no evidence of burning over millennial time scales (Fig. 5). Higuera et al. (2009) hypothesized that a lack of burnable biomass, along with overall cooler summer climate, limited fire occurrence in ancient herb tundra in the south-central Brooks Range, and that similar limits likely apply to the North Slope. In both scenarios, cool summers would not only limit opportunities for fuel drying but also convection, and thus lightning ignition. In the case of the North Slope, increased warmth (e.g., Hinzman et al. 2005), increased vegetation growth (e.g., Tape et al. 2006, Bhatt et al. 2010), and possibly increased lightning density (Hu et al. 2010) may be contributing to an increased fire danger across this region (Jones et al. 2009, Hu et al. 2010). In comparison to more frequent burning in the Noatak study area and ancient shrub tundra, our records suggest that fire on the North Slope is limited by low vegetation abundance (i.e., fuel loads), high fuel or soil moisture due to seasonal climate (including snow cover) or permafrost, or lightning ignitions. All three possibilities should be explored relative to more flammable tundra ecosystems.

Species composition may be more important for determining the nature of fire events, rather than fire occurrence, for example by influencing biomass consumption per fire. Despite similar fire frequencies among down-valley sites and the ancient shrub tundra, charcoal peak magnitudes at down-valley sites were lower than in the ancient shrub tundra. Less charcoal production per fire, on average, suggests that the amount of biomass burned per fire was likewise lower. In contrast, despite lower fire frequencies at Little Isac Lake compared to ancient shrub tundra, charcoal peak magnitudes were similar among these sites (Fig. 5), suggesting a similar amount of biomass burned per fire, on average. Both patterns are consistent with the greater dominance of shrub birch in ancient shrub tundra (Anderson et al. 1994, Higuera et al. 2008), and the higher shrub tundra abundance (Fig. 1) and birch pollen percentages over the past two millennia (Fig. 3) at Little Isac Lake compared to the other Noatak sites. These general patterns suggest that the impacts of future tundra burning, such as fire severity and associated carbon release, will likely vary among different tundra vegetation types.

Implications for land management and Arctic environmental change

The fire-history records presented here help define the historical range of variability of fire in the Noatak region and help managers comply with federally funded programs that promote informed management decisions. Our charcoal-based FRIs are generally shorter than previously estimated mean point-wise fire return intervals (also termed "fire cycles") based on 30–60 years of observation. For example, our estimated mean FRI of 142 years (95% CI = 115–174) for down-valley sites is slightly shorter than previously estimated fire cycles of 175–193 years (Kobuk Ridges and Valleys Ecoregion; Kasischke et al. 2002, Joly et al. 2010*b*). Mean FRI estimates of 263 years (95% CI = 175–374) for up-valley sites are shorter than the 480-yr estimated fire cycles by Gabriel and Tande (1983; Table 2). In both cases, differences likely reflect the difficulty in estimating fire cycles longer than 150 years with a 30-yr data set (Baker 2009) or in determining the precise spatial area used to define short-term fire cycles. Alternatively, differences could reflect a real but undetectable decrease in fire activity, or an overestimation of point-wise fire occurrence from the charcoal records. The latter is unlikely, given that sediment-charcoal records combine fires into fire events within 15-yr intervals.

Through impacts on vegetation, tundra burning also influences wildlife, including economically and culturally valuable caribou and reindeer populations (e.g., Vors and Boyce 2009). For example, fire reduces the long-term availability of fruticose lichen, a key source of winter forage for the Western Arctic Caribou Herd (WAH), which in turn is an important subsistence resource for native Alaskans (Joly et al. 2010*b*). While the majority of fires in western Alaska receive limited suppression (e.g., ~80% of Western Arctic National Parklands, Fire Management Plan for Western Arctic Parklands, National Park Service, 2004, revised 2009; *available online*),[11] land managers have considered increased suppression as a way to protect winter range of the WAH (Joly et al. 2007). Although the Noatak study area is only a portion of this herd's range, our results indicate that if caribou utilized this area over the past 2000 years, then they have successfully coexisted with relatively frequent burning. Landscape patterns of burned and unburned areas were not resolved in detail in this study, but the spatial variability between down- and up-valley sites implies that similar spatial variability elsewhere may have provided enough unburned, and thus suitable, habitat for the WAH in the past.

Across the Arctic, tundra burning and possible shifts in fire regimes are increasingly recognized as important factors affecting ongoing and future environmental change, given fire's impact on a range of biophysical and biogeochemical processes (e.g., Shur and Jorgenson 2007, Wookey et al. 2009, Mack et al. 2011). Our summary of tundra fire history from Alaska (Fig. 5) highlights a biome that can sustain a wide range of burning, with median FRIs from as low as 113 years to more than 5000 years (Table 2, Fig. 5). Despite these widely varying fire histories, it is striking that species composition among Alaskan tundra regions is largely similar across graminoid, deciduous, and evergreen taxa (Walker et al. 2005). If recent tundra burning is truly a harbinger of future environmental change (Jones et al. 2009), then the diverse history of fire across Arctic Alaska suggests that the species makeup of tundra vegetation will be resilient, even to relatively frequent burning. Indeed, preliminary measurements suggest that vegetation recovery within the 2007 Anaktuvuk River Fire, unprecedented in at least 5000 years (Hu et al. 2010), has been surprisingly rapid, due in part to high survival from species like *Eriophorum vaginatum* (cotton grass; M. S. Bret-Harte, R. R. Jandt, D. A. Yokel, P. M. Ray, E. A. Miller, M. C. Mack, and G. R. Shaver, *unpublished data;* P. E. Higuera, *personal observation*). This pattern is consistent with postfire succession in other tundra regions of Alaska, where vegetation is dominated by sedges and grasses in the first several postfire years (Wein and Bliss 1973, Racine et al. 2004, Jandt et al. 2008). The more important impacts of tundra burning on vegetation will likely be through changes in relative species abundance that emerge over many years to decades. For example, fire may facilitate increased shrub abundance in arctic and subarctic tundra (e.g., de Groot and Wein 1999, Racine et al. 2006, Joly et al. 2010*a*, Lantz et al. 2010*a*), implying that future burning could amplify the direct effects of climatic warming on tundra plant communities (Euskirchen et al. 2009).

Fire is also a widely recognized mechanism that could lead to the rapid release of greenhouse gases into the atmosphere, through direct and indirect emission of carbon dioxide (McGuire et al. 2009) and methane (Osterkamp 2007, Schuur et al. 2009). Although the role of tundra fires in the carbon cycle is poorly understood (Sitch et al. 2007), recent evidence suggests that they can lead to the immediate loss of significant carbon stocks (Mack et al. 2011). Longer-term impacts of tundra burning on carbon and methane emission could come through accelerated permafrost thawing (Shur and Jorgenson 2007), which in turn facilitates the breakdown of ancient organic matter. The wide range of past tundra burning highlighted here suggests that the role of tundra fires in shaping these biophysical and biogeochemical processes should likewise vary. For example, long-term carbon accumulation rates in soil and organic matter could vary significantly between systems that

[11] ⟨http://www.nps.gov/akso/fire/policy/docs/WesternArcticParklands-FireManagementPlan-2009.pdf⟩

have burned frequently over the past several millennia (e.g., the Noatak study area) compared to those that have not burned for millennia (e.g., the Anaktuvuk River Fire region). Likewise, if disturbance by fire is a key determinant of permafrost stability (Shur and Jorgenson 2007), then permafrost formation and thaw should vary among systems that have experienced frequent burning in the past relative to those that have not.

The great variety of fire histories in the tundra biome highlighted here cautions against inferring any single link between tundra burning and biophysical and biogeochemical processes in Arctic systems. The historical frequency of fire in a system likely plays an important role in determining the sensitivity and resilience of that system to changing climate and possibly fire regimes. Ongoing research linking fire with other biological and physical processes can use the diverse history of tundra fire regimes highlighted here to help test different hypotheses and examine patterns across a range of fire, climate, and vegetation gradients.

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APPENDIX A

Supplementary study site information, including photographs, bathometric maps and coring locations, and modern vegetation and climate for Raven, Uchugrak, Poktovik, and Little Isac lakes (*Ecological Archives* A021-144-A1).

APPENDIX B

Supplementary results: radiometric dates (^{14}C and ^{210}Pb) and age–depth models, and supplementary pollen and charcoal records for Raven, Uchugrak, Poktovik, and Little Isac lakes (*Ecological Archives* A021-144-A2).

SUPPLEMENT 1

Google Earth file (.kmz) with location of lakes used and referenced in this study, and polygons of recorded fires in Alaska from 1950 to 2007 (*Ecological Archives* A021-144-S1).

SUPPLEMENT 2

Age–depth models for Raven, Uchugrak, Poktovik, and Little Isac lakes, including Monte Carlo bootstrapped confidence intervals (*Ecological Archives* A021-144-S2).

SUPPLEMENT 3

Raw and analyzed charcoal data for the past 3000 years for Raven, Uchugrak, Poktovik, and Little Isac lakes (*Ecological Archives* A021-144-S3).

SUPPLEMENT 4

Pollen counts for the past 3000 years for Raven, Uchugrak, Poktovik, and Little Isac lakes (*Ecological Archives* A021-144-S4).