# Carbon emissions and radiative forcings from tundra wildfires in the Yukon-Kuskokwim River Delta, Alaska

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**Abstract**

Tundra environments are experiencing elevated levels of wildfire, and the frequency is expected to keep increasing due to rapid climate change in the Arctic. Tundra wildfires can release globally significant amounts of greenhouse gasses that influence the Earth's radiative balance. Here we develop a novel method for estimating carbon loss and the resulting radiative forcings of gaseous and aerosol emissions from 2015 tundra wildfires in the Yukon-Kuskokwim Delta (YKD), Alaska. We paired burn depth measurements using two vegetative reference points that survived the fire event —*Sphagnum fuscum* and *Dicranum spp.*— with measurements of local organic matter and soil carbon properties to estimate total ecosystem organic matter and carbon loss. We used remotely-sensed data on fire severity from Landsat 8 to scale our measured losses to the entire fire-affected area, with an estimated total loss of 2.04 Tg of organic matter and 0.91 Tg of carbon, and an average loss of 3.76 kg m$^{-2}$ of organic matter and 1.68 kg m$^{-2}$ of carbon in the 2015 YKD wildfires. To demonstrate the impact of these fires on Earth's radiation budget, we developed a simple but comprehensive framework to estimate the radiative forcing from Arctic wildfires. We synthesized existing research on the lifetime and radiative forcings of gaseous and aerosol emissions of $CO_2$, $N_2O$, $CH_4$, $O_3$ and its precursors, and fire aerosols. The model shows a net positive cumulative mean radiative forcing of 3.67 W m$^{-2}$ using RCP 4.5 and 3.37 W m$^{-2}$ using RCP 8.5 at 80 years post-fire, which was dominated by $CO_2$ emissions. Our results highlight the climate impact of tundra wildfires, which positively reinforce climate warming and increased fire frequency through the radiative forcings of their gaseous emissions.

**1 Introduction**

The Arctic region is characterized by permafrost soils with low rates of decomposition and high carbon content from millennia of positive net ecosystem production (NEP; Lindgren et al., 2018). As a result, there is more than twice as much carbon stored in permafrost soils as there is in the atmosphere, including roughly 1,035 ± 150 Pg of carbon in the top three meters of soil (Schuur et al., 2015). Surface air temperatures in the Arctic have been increasing more than twice as fast as the global average since the mid-20th century (Cohen et al., 2018) and are expected to continue increasing with more variable precipitation, hence more frequently pairing hot and dry conditions (Hu et al., 2015). Changing environmental conditions will place some of this large carbon stock at risk for release into the atmosphere through increased biological activity and wildfire (Natali et al., 2019, Natali et al., 2015, Rocha et al., 2011a, Hu et al., 2010).

62    Fires are an increasingly important component of tundra carbon cycling. Fires are becoming more frequent

63  in Arctic systems due to increasing occurrences of hot and dry conditions coupled with more lightning ignitions

64  (Chen et al., 2021, Bieniek et al., 2020, Veraverbeke et al., 2017b). For example, the 2007 Anaktuvuk River

65  megafire on the North Slope of Alaska occurred during an especially hot and dry period and released an amount of

66  carbon similar in magnitude to annual sequestration across the entire tundra biome (Mack et al., 2011). Continued

67  and potentially accelerated rates of warming are expected to further increase the frequency of tundra wildfires,

68  thereby releasing significant amounts of carbon and altering the net carbon balance of the tundra biome (Hu et al.,

69  2015).

70    A recent body of literature indicates the potential to estimate wildfire carbon emissions in boreal forests by

71  linking geospatial predictors, most prominently satellite-derived estimates of fire severity and extent, with *in situ*

72  measurements of carbon loss (e.g. Dieleman et al., 2020, Walker et al., 2018b, Veraverbeke et al., 2015b, and

73  Rogers et al., 2014). Despite increasing incidence of tundra wildfires, these studies have primarily focused on boreal

74  forests, and few estimates are available for tundra (Mack et al., 2011). Whereas satellite-derived fire extent and

75  severity is widely available, representative *in situ* measurements of tundra wildfires are not. *In situ* measurements of

76  fire effects on tundra organic soils are challenging due to a lack of reference points (e.g., adventitious roots) that

77  survive the fire, which are used to estimate pre-fire organic matter depth and content (Walker et al., 2018a, Rogers et

78  al., 2014, Boby et al., 2010). Measurements of organic matter loss must also be region-specific because of

79  differences in vegetation and soil properties (Walker et al., 2020c, Mack et al., 2011).

80    Although carbon loss estimates from Arctic wildfires are important for understanding the impacts of

81  climate change on carbon budgets, radiative forcings from wildfire gaseous and aerosol emissions are needed to

82  properly gauge the impact on the Earth's atmosphere and climate (Huang et al., 2016, Ward et al., 2012, Randerson

83  et al., 2006). Radiative forcings from wildfires depend spatially on fire severity and atmospheric conditions and

84  temporally on changing atmospheric background concentrations in the months, years, and decades following the fire

85  (Huang et al., 2016, Joos et al., 2013). To date, radiative forcings from high-latitude wildfires have been estimated

86  for particular boreal forests (Huang et al., 2016, O'Halloran et al., 2012, Randerson et al., 2006) and within global

87  models (Ward et al., 2012), but not for tundra ecosystems.

88    Here we take a two-step approach to assess the impact of tundra wildfires on carbon budgets and climate.

89  We first developed a method for measuring carbon loss *in situ* in tundra ecosystems, particularly for the 2015 fire

90   season in the Yukon-Kuskokwim Delta, AK. We tested the agreement of multiple pre-fire reference points for

91   estimating burn depth in the field and combined these estimates with laboratory-measured organic matter and carbon

92   fractions to estimate emissions. We then used 30m remotely-sensed fire severity data (differenced Normalized Burn

93   Ratio, or dNBR; Key and Benson, 2006) from Landsat 8 to scale our measurements to the entire fire area.  Finally,

94   we estimated the long-term radiative forcings of the fire season's gaseous and particulate emissions, including long-

95   lived greenhouse gasses (GHGs), ozone, ozone precursors, and aerosols, using a variety of published algorithms and

96   arctic-specific parameters when available.

97   **2 Materials and Methods**

98   **2.1 Study Area**

99        Field measurements were collected in the summer of 2019 in a burn scar from the 2015 fire season in the

100  Yukon-Kuskokwim Delta, AK (YKD; Fig. 1).  The burn scar we sampled was adjacent to an uninhabited but

101  regularly accessed Arctic field research outpost in the YKD established in 2016.  Base camp was situated at 61.2632

102  ºN, 163.2458 ºW, approximately 95 kilometers northwest of Bethel, AK, accessible by float plane and helicopter.

103  Sampling was done within about an 11-kilometer radius of the base camp (Fig. 1c).  Measurements from the field

104  were scaled to all fire scars in the YKD that burned in 2015 (Fig. 1b). We used fire perimeters from the Alaska

105  Large Fire Database (ALFD; Kasischke et al., 2002).

106

107
108 **Figure 1.** (a) The location of the YKD within Alaska, (b) the areas of fires in 2015 within the YKD, and (c) the
109 locations of our vegetative reference point height transects in burned and unburned areas within about an 11 km
110 radius from base camp. Fire perimeters were derived from the Alaska Large Fire Database (Kasischke et al., 2002).
111 Sources: Esri, DigitalGlobe, GeoEye, i-cubed, USDA FSA, USGS, AEX, Getmapping, Aerogrid, IGN, IGP,
112 swisstopo, and the GIS User Community.

113
114     The YKD contains tussock-sedge, dwarf-shrub, and lichen and moss tundra communities over ice-rich

115 permafrost peat plateaus (Raynolds et al., 2005).  These peat plateaus were the primary land component burned in

116 the wildfires and were separated by fens, bogs, and open water.  Within our sample area, lichen was the dominant

117 vegetation type (Frost et al., 2020), shrubs were interspersed across the landscape, and tussocks were rare.  Soil is

118 predominantly organic (Jorgenson et al., 2000), which has accumulated over sand deposits from the Pleistocene

119 period (Shaw, 1998).

120 **2.2 Fire severity**

121     Fire perimeters in the YKD in 2015 were extracted from the ALFD.  We used a remotely sensed metric of

122 fire severity from Landsat 8 Tier 1 surface reflectance imagery: dNBR (Eq. 1), which is based on the Normalized

123  Burn Ratio (NBR; Eq. 2; Key and Benson, 2006). Clouds, cloud shadows, and snow were masked from all images

124  using the provided pixel quality attributes generated from the CFMASK algorithm (Foga et al., 2017). Imagery for

125  this index was acquired as the mean Landsat 8 composite between June 1st and August 31st for one year pre- and one

126  year post-fire, and for all fire scars. All remotely sensed variables were extracted in Google Earth Engine (Gorelick

127  et al., 2017). Where in Eq. (2) NIR is near-infrared (Band 5) and SWIR is shortwave infrared (Band 7).

128  $dNBR = (NBR_{prefire} - NBR_{postfire}) \times 1000,$ (1)

129  $NBR = \frac{(NIR - SWIR)}{(SWIR + NIR)},$ (2)

130  We compared dNBR from the YKD 2015 fire season to all fires in boreal and tundra Alaska to place the

131  measured dNBR in context. Using fire records from the ALFD, dNBR was acquired for all available fire perimeters

132  in Alaska. Differenced NBR was calculated in the same way as described before, but with the exception that

133  Landsat 4, 5, 7, and 8 were used in the pre- and post-fire composites. Due to differences in spectral bands, Landsat 8

134  was corrected to match Landsat 7 using a regression technique (Roy et al., 2016). The first fire year we detected

135  dNBR was in 1989 and the most recent was in 2019. Boreal and tundra extents were defined using the United States

136  Environmental Protection Agency ecoregions, with tundra as level one class 2.0 and boreal as level one classes 3.0

137  and 6.0 (U.S. Environmental Protection Agency, 2010).

138  **2.3 Vegetative marker height measurements**

139  We measured the height of vegetative reference points above the surface in both unburned and burned

140  areas. Three main dominant surface vegetative reference points were available: *Sphagnum fuscum, Dicranum spp.,*

141  and *Eriophorum vaginatum* (tussocks; similar to the methods of Mack et al. (2011)). *S. fuscum* moss appeared in

142  large mats. *Dicranum* moss appeared in small dense patches. All vegetative reference points were particularly

143  conspicuous in burned areas as they were elevated above the burned surface (Fig. 2). In the burned area, we

144  measured the distance from the soil surface to top of the living parts of the vegetative reference points, which we

145  assumed to indicate full survival in the fire event. We did not measure dead remnants of vegetative reference points,

146  because we expect that these may not represent the actual pre-fire vegetation height.

147
148  **Figure 2.** An example burn depth measurement from the surviving top of a *Dicranum spp.* moss patch (red arrow;
149  pink string) to the top of the soil (blue arrow) in a burned transect. Transect tape was used to measure distance
150  between the moss patches.
151
152          We measured the average height along a transect between two comparable reference points. At the highest

153  living point on each moss patch, or to the top of the corms at the base of a tussock, we inserted a nail (red arrow,

154  Fig. 2). We ran a taut piece of string between the two nails (Fig. A1). In between the edges of the moss patch or

155  tussock pair (Fig. A1), we took height measurements vertically from the string to the ground, which was soil in the

156  burned areas (blue arrow, Fig. 2) or a dense vegetative surface in the unburned areas (Fig. A2). In between the

157  starting and ending point measurements on the transect we recorded height every 25 centimeters (Figs. A2 and A3).

158  The starting and ending points were never greater than 50 centimeters from the nail. Latitude and longitude were

159  recorded at all transects with an accuracy of three meters (Fig. 1) for ground-truthing remotely sensed burn severity

160  categories.

161          One potential source of error is post-fire subsidence between the vegetative reference point pairs that may

162  bias the relative height of a vegetative marker above the surface (Jones et al., 2015). We minimized this potential

163  bias by maintaining transect lengths less than four meters, and predominantly one to two meters in burned areas.

164  We also visually assessed the area between each vegetative marker for signs of subsidence, including cracked soil or

165     large elevation differences, and chose not to sample these areas. Measuring from the top of one vegetative marker to

166     the top of another marker controlled for slopes and larger landscape elevation features because the string largely

167     mimicked the angle of the landscape.

168          We selected transects for height measurements opportunistically in the burned and unburned areas.

169     However, we tried to maximize spatial separation between transects and target visually identifiable areas of varying

170     burn severity. We corrected for spatial autocorrelation biases between transect averages in our statistical analyses

171     (see below). In unburned areas we maximized the number of transects measured for each vegetative reference point,

172     *Sphagnum* (n = 38), *Dicranum* (n = 40), and *Eriophorum* (n = 19).

173          Each burn depth measurement was associated with a dNBR value based on the 90 m pixel, the mean of

174     underlying 30 m pixels, containing its centroid. Burn depth measurements that came from the lowest third, middle

175     third, and upper third of dNBR values were considered to be in low severity/unburned, moderate severity, and high

176     severity areas, respectively, and a roughly even numbers of transects were sampled in each burn severity category

177     for *Sphagnum* (n = 56: high n = 20, mod n = 17, low/unburned n = 19), *Dicranum* (n = 54: high n = 18, mod n = 17,

178     low/unburned n = 19), and *Eriophorum* (n = 24: high n = 7, mod n = 11, low/unburned n = 6). Unburned patches of

179     tundra were sometimes included in low severity pixels within the burn scar, which is why these severity classes are

180     combined.

181     **2.4 Organic matter and carbon pool measurements**

182          We calculated the total ecosystem organic matter and carbon pool sizes in unburned areas surrounding the

183     fire scar. We took vegetation and soil samples together in cores using a hand drill and hollow metal drill bit that was

184     30 centimeters in length and six centimeters in diameter. Cores were extracted at three points (start, middle, and

185     end) along transects between like vegetative reference points. We selected four sites of less than half a hectare

186     surrounding areas where we took our unburned vegetative reference point height measurements. These sites were

187     selected opportunistically to ensure the presence of an appropriate number of each of the three vegetative reference

188     points. In each site, we extracted cores from three transects per vegetative reference point pair, which totaled 27 soil

189     cores per site. In one site we found only two tussock pairs, so our total number of samples was 105. We chose

190     unburned sites separated by at least one kilometer to control for heterogeneity in local soil and vegetation

191     characteristics.

192       We measured the height of the live vegetation layer and fibric soil horizons for each core.  Vegetation

193    layers and soil horizons were identified visually by soil density, texture, color, and identifiable plant parts.  Each

194    core was separated into vegetation and fibric soil layers that were homogenized, weighed, and subsampled in the

195    field, and subsequently stored frozen until analyzed.  In the lab, each vegetation sample and approximately 15 grams

196    of each fibric sample were dried for 48 hours at 60 ℃ and weighed to determine soil water content.  We used the

197    proportion of dry mass to wet mass to estimate the dry weight of the field sample, which was used to determine bulk

198    density (g dry mass cm$^{-3}$).  Soil samples were combusted for five hours at 450 ℃ to determine organic matter

199    content and analyzed for percent carbon using an Elementar Vario Max CN analyzer at the Woodwell Climate

200    Research Center, Falmouth, MA.  We assumed the dry mass of the vegetation layer was fully organic matter.  Using

201    the average across all herbaceous plant organs, we assumed the carbon content of the vegetation layer was 43.04%

202    from Ma et al. (2018).  To calculate organic matter and carbon pools (kg m$^{-2}$) for each organic soil sample, we

203    multiplied bulk density by the height of the layer and percent organic matter or percent carbon, respectively.  These

204    values were normalized for each sample to the average depth of the vegetation layer, which was seven centimeters,

205    and 10 centimeters in the fibric horizon.

206    **2.5 Calculating final estimates**

207       We calculated organic matter combustion and carbon loss across the fire area using burn depth derived

208    from the moss reference points. We excluded estimates derived from tussock measurements, because our sample

209    size of tussock-based measurements was substantially lower than the moss vegetative reference points due to their

210    infrequent occurrence in our study area (Frost et al., 2020).  Moreover, burn depth estimates from tussocks

211    correlated negatively with remotely sensed fire severity (Fig. A4).   This negative correlation with fire severity may

212    be attributable to mechanisms that are untestable with our sample size and study design, including altered burn

213    dynamics in close proximity to tussocks.  Future work could clarify the mechanism behind this discrepancy.

214    Excluding tussock measurements required reassigning dNBR values evenly as before across only *Dicranum* (n = 54:

215    high n = 18, mod n = 18, low/unburned n = 18) and *Sphagnum* (n = 56: high n = 20, mod n = 18, low/unburned n =

216    18) reference points.  The highest dNBR values in the low severity/unburned and moderate severity categories

217    became the threshold values for low severity/unburned to moderate severity and moderate severity to high severity

218    burn, respectively.  We split the fire area, based on these thresholds, into the three categories for burn severity.

219    There were 13,001 hectares of low severity burn, potentially including unburned patches, 8,516 hectares of moderate

220 severity burn, and 32,637 hectares of high severity burn, which summed to 54,154 hectares of total fire area (Fig.

221 A5).

222 For final analysis of organic matter and carbon loss, we averaged the height of *Sphagnum* and *Dicranum*

223 reference points in the unburned areas to determine one height for each reference point, since their respective

224 heights were statistically different (p < 0.001).  Burn depth was calculated at each transect in the burned areas by

225 subtracting the standard unburned height from the transect average height.  We also averaged the dry organic matter

226 and carbon pool within the vegetative layer and fibric horizon separately across *Sphagnum* and *Dicranum* reference

227 points to determine one pool size for our sampling area.  Average organic matter or carbon loss (kg m$^{-2}$) was

228 estimated as the product of the organic matter or carbon concentrations and depth of the vegetation and soil that

229 burned at each burn depth transect.  Average loss at each transect was the sum of loss from each horizon (see Fig.

230 A6 for schematic of organic matter and carbon loss calculation per transect). For our final calculation of total dry

231 organic matter and carbon loss, we averaged the organic matter and carbon lost across transects for *Sphagnum* and

232 *Dicranum* reference points within each fire severity category.  Over the landscape, total carbon or organic matter

233 released was calculated by multiplying average carbon or organic matter loss for a given burn severity category by

234 the total burned area for that category.  Total carbon or organic matter released was then calculated as the sum from

235 all three burn severity categories.

236 **2.6 Radiocarbon**

237 We used radiocarbon dating to confirm our burn depth measurements by comparing the radiocarbon age of

238 *Sphagnum* macrofossils on the burned soil surface with the radiocarbon age of *Sphagnum* macrofossils at different

239 depths of the unburned soil profile, similar to the methods of Mack et al. (2011).  In the field we extracted one

240 unburned core and three burned cores of the same dimensions as our primary soil cores, described above.  We

241 extracted a one-centimeter thick subsample of the cores every five centimeters in depth and froze the samples in a

242 sealed plastic bag until lab analysis.  To construct an age profile, we analyzed unburned cores at depths of five to 20

243 centimeters for radiocarbon age.  We assumed that a depth of zero centimeters had carbon from the time of harvest

244 in 2019. In the burned cores, we analyzed the zero-centimeter depth (i.e., surface) for radiocarbon.

245 In the lab, we visually identified *Sphagnum* branches in each of the subsamples being tested and rinsed

246 them with deionized water.  The branches were then dried for 48 hours at 60 ℃, and roughly five milligrams of dry

247 branch matter per sample was sent to the W.M. Keck Carbon Cycle Accelerator Mass Spectrometry Laboratory at

248    the University of California Irvine for the $^{14}$C content and $\delta^{13}$C. Results for $^{14}$C analysis are reported as the fraction

249    modern, which is defined as the ratio of the sample's carbon isotope ratio to that of a standard (Reimer, 2004).

250    Fraction modern values are calibrated to calendar years using OxCal online 4.4.1 (Ramsey, 2009), with the

251    appropriate modern era bomb curve taken from Hua et al. (2013). Due to the shape of the bomb curve, there are at

252    least two possible calendar years in which the carbon in the sample was fixed. However, we did not use the oldest

253    age for unburned samples with multiple calendar ages because the age of these samples must increase with depth

254    (Walker et al., 2019, Mack et al., 2011), which is not supported when considering the oldest calendar ages.

255    Furthermore, we were confident the burned samples were from the younger age because they were from visually

256    shallow burn depths. We compared the calendar age of carbon at the burned soil surface to age by depth profile

257    taken from the unburned core.

**2.7 Radiative forcings model**

259        We created a temporally-explicit model of radiative forcings for gaseous and aerosol emissions of tundra

260    wildfires and used it to compute the radiative forcings per unit burned area. The radiative forcings model was

261    driven using the average amount of organic matter lost across fire-wide burn severity classes and vegetative

262    reference points. Computing the radiative forcing of gaseous and aerosol emissions has been done for boreal fires

263    (e.g., Huang et al. 2016, O'Halloran et al., 2012, Randerson et al., 2006), but has yet to be applied to tundra systems.

264    Our model included the long-lived GHG species $CO_2$, $CH_4$, and $N_2O$, as well as short-lived climate forcers,

265    tropospheric $O_3$, $O_3$ precursors, and aerosols. Ozone precursors include $NO_x$, non-methanogenic volatile organic

266    carbons (NMVOCs), and CO.

267        We first used emissions factors from Akagi et al. (2011) to calculate the mass of gaseous and aerosol

268    emissions from our estimated organic matter losses. Emissions factors have not been previously defined for tundra

269    burning. Given that boreal forest, whose definition for emission factors includes organic soils, peat, and woody

270    vegetation, is likely the closest ecosystem type in terms of fuel properties to tundra in Akagi et al. (2011), we

271    employed these emission factors. However, we note a possible overestimate of relative contribution from woody

272    vegetation emissions in these numbers due to the relative lack of woody vegetation on the tundra landscape.

273    Furthermore, concentrations of gasses released from combustion of the same biomass type vary based on the

274    measurement technique, such as from differences in chemical mixing in laboratory, airborne, and ground-based field

275    methods (Akagi et al., 2011). Since Akagi et al. (2011) provides estimates of each emissions factor based on a

276    combination of values from previous studies, the emissions factors may capture uncertainty associated with the

277    variable methods of these source studies.

278        Once we estimated the mass of each gaseous emission, we calculated the concentration of the gas

279    remaining in the atmosphere and its radiative forcing each year after the fire season by synthesizing existing models

280    and research on the lifetimes and radiative forcings of these gaseous emissions. Our calculations of radiative forcing

281    were dependent on the future ambient concentration of GHGs in the atmosphere. As a result, we calculated the

282    radiative forcings of gaseous emissions for three scenarios: historic, representative concentration pathway (RCP)

283    4.5, and RCP 8.5. The historic scenario assumes the ambient concentration of GHGs remains constant in the

284    atmosphere after the fire year. Future atmospheric concentrations for each RCP were taken from Meinshausen et al.

285    (2011). The radiative forcing for each gaseous emission per year was calculated separately and then summed across

286    forcing agent. Since $O_3$ precursors and aerosols had the most uncertain lifetimes and radiative forcings (Bond et al.,

287    2011, Quinn et al., 2008), we calculated the total radiative forcing of the emissions with and without them. A

288    flowchart of our general methodology for the entire radiative forcings calculation is presented in the appendix (Fig.

289    A7).

290    **2.8 Methane and Nitrous Oxide**

291        Our radiative forcing calculations for $CH_4$ and $N_2O$ were based on Ward et al. (2012). For both gasses, we

292    assumed a pulse emission and calculated the concentration remaining in each year after the fires using a simple box

293    model with one outflow as shown in Eq. (3), where $C_o$ is the initial pulse concentration of the gas, t is the number of

294    years after the fire event, and L is the lifetime of the gas. The initial pulse concentration of both gasses was

295    calculated by converting the mass emitted to a volume as a molar fraction of the atmosphere.

296    $C = C_o e^{\frac{-t}{L}},$                (3)

297    $L = L_o \left(\frac{E}{E_t}\right)^{-0.05},$           (4)

298

299        We used the atmospheric lifetimes for both gasses reported in Myhre et al. (2013). However, we calculated

300    a decreased lifetime of $N_2O$ using Eq. (4) from Ward et al. (2012), as its ambient concentration changed in future

301    emissions scenarios, where $L_o$ is the initial lifetime of $N_2O$ from Myhre et al. (2013), E is the ambient concentration

302    of $N_2O$ in the year the fires burned, and $E_t$ is the ambient concentration of $N_2O$ at each year after the fire event based

303    on the future emissions scenario (Meinshausen et al., 2011). We did not account for the effect of the changing

304    atmospheric concentration on the lifetime of $N_2O$ due to the pulse emission itself. We then calculated the

305    perturbation concentrations of $CH_4$ and $N_2O$ in the atmosphere each year after the fire event as the sum of the

306    remaining pulse emissions and the ambient concentrations and used Eqs. (5) and (6) from Ward et al. (2012) to

307    estimate the radiative forcing of both gasses. The radiative forcing of the gas at each year was RF, M was the

308    perturbation concentration of the gas whose radiative forcing was being calculated, and $M_o$ and $N_o$ were the ambient

309    concentrations of both gasses depending on which gas's radiative forcing was being calculated.

310    $$RF = 0.036(\sqrt{M} - \sqrt{M_o}) - [f(M, N_o) - f(M_o, N_o)], \tag{5}$$
311    $$f(M,N) = 0.47 ln[1 + 2.01 \times 10^{-5}(MN)^{0.75} + 5.31 \times 10^{-15} M(MN)^{1.52}], \tag{6}$$
312
313    **2.9 Carbon Dioxide**

314        Carbon dioxide has a highly variable lifetime under different future emissions scenarios depending on the

315    strength of ocean and land sinks. To account for this variation, we used impulse response functions (IRFs) from

316    Joos et al. (2013), which represent the fraction of a pulse of $CO_2$ remaining in the atmosphere at each year after the

317    pulse for each scenario. For each year post-fire, we multiplied the relevant IRF function by the concentration of

318    $CO_2$ initially released to estimate the amount of $CO_2$ remaining in the atmosphere. Similar to $N_2O$ and $CH_4$, the

319    initial concentration of $CO_2$ was calculated by converting the mass emitted to a volume as a molar fraction of the

320    atmosphere. The radiative forcing for $CO_2$ was calculated using Eq. (7) taken from Myhre et al. (1998), where C is

321    the perturbation concentration and $C_o$ is the ambient concentration of the gas in each year.

322    $$RF = 5.35 \, ln\left(\frac{C_o + C}{C_o}\right), \tag{7}$$
323    **2.10 Tropospheric Ozone**

324        Tropospheric $O_3$ is created photochemically in the smoke plumes of wildfires by the combination of

325    sunlight and $O_3$ precursor gasses, and its creation can be expressed as a function of time and the amount of carbon

326    monoxide released (Jaffe and Wigder, 2012). We used data compiled in Jaffe and Widger (2012), who provided a

327    synthesis of $O_3$ and its relationship with CO for fire plumes sampled in the Arctic at various times after a fire to

328    calculate a mean plume lifetime and derive a linear relation between plume age and the ratio of $O_3$ to CO. Because

329    the average plume age sampled was five days, we assumed $O_3$ was created for five days after the fires. The

330    concentration of $O_3$ on each of those five days was calculated by multiplying the initial concentration of CO released

331    from the fires, calculated by converting the mass emitted to a volume as a molar fraction of the atmosphere, by the

332    ratio of $O_3$ to CO we estimated for that day. After five days, we assumed no more $O_3$ was created, and we

333    calculated the remaining amount of $O_3$ after day five using the box model approach from Eq. (3). The concentration

334    of $O_3$ on day five was $C_o$, and L was the lifetime of $O_3$ taken from Myhre et al. (2013). We converted the

335 concentration of $O_3$ to Dobson units (DUs) and converted from DUs to radiative forcing using a conversion factor

336 from Myhre et al. (2013) each day after the fire event.  The radiative forcing in each year was the mean radiative

337 forcing of all the days within that year, although the annual forcing was negligible after the first year.

338 **2.11 Ozone Precursors and Aerosols**

339 We used a method based on global warming potential (GWP), similar to Huang et al. (2016), to calculate

340 the radiative forcing of the ozone precursors and the direct aerosol effect.  GWP is defined as the ratio of the time-

341 integrated radiative forcing of a pulse emission of a gaseous species to the time-integrated radiative forcing of an

342 equivalent emission mass of $CO_2$ (Joos et al., 2013).  GWP is typically calculated for 20- and 100-year time

343 horizons.  Here, we use GWPs to determine the warming effect of the precursor and aerosol emissions relative to

344 $CH_4$ at the 20- and 100-year time horizons, as shown in Eq. (8).  We derived the radiative forcings for ozone

345 precursors based on $CH_4$ because the radiative forcing of ozone precursors is through their effect on methane over

346 the long-term (Collins et al., 2013).  To remain consistent with our methodology for ozone precursors, we derive the

347 radiative forcings for aerosols from $CH_4$ as well.  Furthermore, the radiative forcing of $CH_4$ has the simplest

348 derivation in our model, so we assume it is the strongest continuous radiative forcing from which to build our

349 continuous model using GWPs.

350 $$R_t = \frac{GWP_{t,x} \times EF_x}{GWP_{t,CH_4} \times EF_{CH_4}},$$  (8)

351 The emissions factor (EF) and GWP are defined for gaseous species x, and the GWP is defined at the time

352 horizon t.  For CO and NMVOCs, we interpolate this relative warming effect ($R_t$) between year one and year 20 by

353 assuming it remains constant.  To interpolate this effect between years 20 and 100, we mirror the shape of methane's

354 cumulative radiative forcing curve over that time interval.  From this curve we calculated the difference between

355 methane's cumulative radiative forcing at years 20 and 100.  We then calculated the fractional decrease of this

356 difference each year between years 20 and 100.  We multiplied those fractional decreases by the difference in $R_t$ at

357 years 20 and 100 for CO and NMVOCs to derive a relative warming effect for the precursors.  Finally, we

358 multiplied $R_t$ by methane's cumulative radiative forcing curve to estimate the cumulative radiative forcing for each

359 precursor every year after the fire event.  We used the global GWPs for CO and NMVOCs from Myhre et al. (2013).

360 Since $NO_x$ has a positive GWP at the 20-year time horizon and a negative GWP at the 100-year time

361 horizon, we chose not to mimic the shape of methane's radiative forcing curve when interpolating between the two

362 time points.  Instead, we calculate an $R_t$ value for $NO_x$ at year one by scaling $R_{20}$ in proportion to the change of

363  methane's cumulative radiative forcing from years one to 20.  Then we interpolated between the $R_t$ values in years

364  one, 20, and 100 using an exponential decay function of the form $ae^{-bx} + C$, whose coefficients were calculated

365  using the "nls" function in R (R Core Team 2020).  We multiplied the $R_t$ value for $NO_x$ at each year by the

366  cumulative radiative forcing of $CH_4$ in that year to obtain a cumulative radiative forcing curve for $NO_x$.  We use the

367  global GWP for $NO_x$ from Myhre et al. (2013).

368       The radiative effect of aerosol emissions happens within a year of the fire event, as fire aerosols are

369  typically removed from the atmosphere via wet and dry deposition within a matter of weeks (Bond et al., 2011,

370  Quinn et al., 2008).  We assumed that the cumulative radiative forcing of aerosols at any year after the fire event

371  would be constant and equal to the radiative forcing of aerosols in the year of the fire event.  Therefore, to calculate

372  the cumulative radiative forcing from black and organic carbon direct effects in every year, we used $R_{20}$ and

373  multiplied this value by the cumulative radiative forcing curve of $CH_4$ at year 20.  Radiative forcings of black and

374  organic carbon were summed to report a single value for aerosols.  We used the GWPs for black and organic carbon

375  estimated for open biomass burning including the cryosphere effect from Bond et al. (2011).  To estimate the

376  indirect aerosol effect, we multiplied the radiative forcing of the direct effect of aerosols each year by the ratio of

377  indirect to all sky direct effect radiative forcing from wildfires defined in Ward et al. (2012).

378  **2.12 Statistical analyses**

379       All statistical analyses were completed in R (R Core Team 2020) using the nlme package (Pinheiro et al.,

380  2020).  For the vegetative reference point heights, we used the nlme package function "gls" to fit a linear model

381  using the generalized least squares method with average burn depth and height above the dense vegetation layer

382  along the transect as the response variable in the burned and unburned areas respectively.  Within the "gls" function,

383  both models were corrected for spatial autocorrelation between transect locations by choosing the model with the

384  lowest Akaike information criterion (AIC) score across five correlation structures, exponential, Gaussian, linear,

385  rational quadratic, and spherical, to be our final model.  In burned areas, the model had vegetative reference point

386  nested within burn severity type, while in unburned areas, we only modeled the effect of vegetative reference point

387  type.  The model with the lowest AIC score for both burned and unburned areas had a rational quadratic correlation

388  structure.  To analyze differences in organic matter and carbon pools, height, and bulk density of the vegetation and

389  fibric soil layers we used the "lme" function, defined in the nlme package, to fit linear mixed effect models with the

390  restricted maximum likelihood method.  These models had the soil or vegetation layer characteristic value for each

391 core as their response variable, vegetative reference point type as a fixed effect, and transect number nested in site as

392 random effects.  Finally, we created linear fit models using the "gls" function with organic matter combusted and

393 carbon lost at each transect in the burned areas as response variables with the same structure as the burn depth

394 model.  These models had the lowest AIC scores with a rational quadratic correlation structure.  An analysis of

395 variance (ANOVA) was used to test for significant differences between groups for all of our models.

396 **3 Results**

397       We found that burn depth measurements were consistent with burn severity classification, and burn depth

398 increased with fire severity (Fig. 3).  Based on a two-factor ANOVA with vegetative reference point nested within

399 burn severity, differences in average transect burn depth were significant ($p < 0.0001$) between burn severity

400 categories.  The ranges of moderate severity burn depth measurements overlapped substantially with the high

401 severity and low severity/unburned measurements.  Moderate severity groups also had the most evenly distributed

402 probability densities across burn depth.  *Dicranum* measurements captured the shallowest and the deepest burn

403 depths (Fig. 3).  Nevertheless, across both vegetative reference points, there was a clear stepwise increase in burn

404 depth across burn severity classifications.

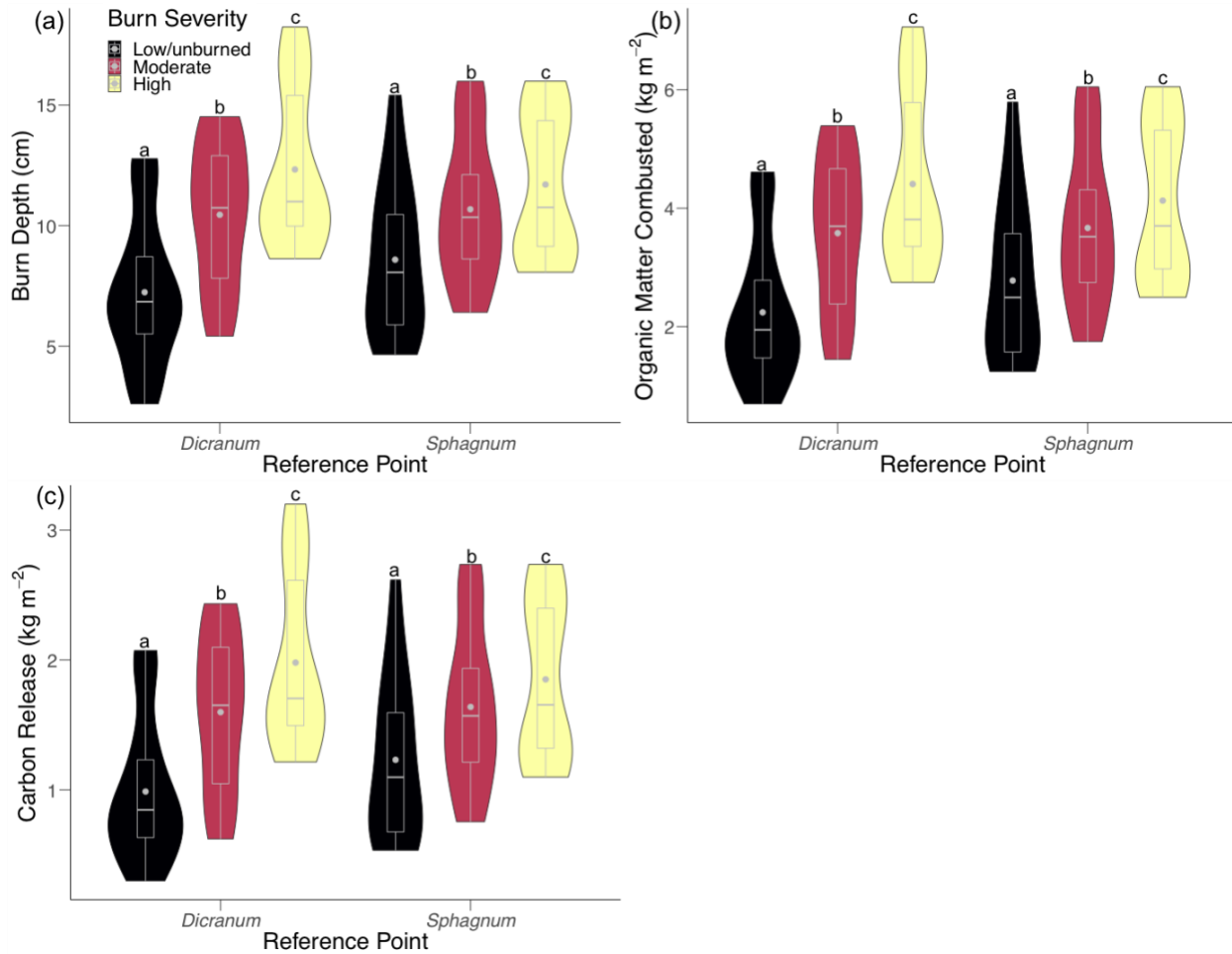



**Figure 3.** Distributions of average (a) burn depth, (b) organic matter combusted and (c) carbon released per unit burned area per transect within each reference point across burn severity levels. Boxes encompass the middle 50% of data, whiskers are the upper and lower quartiles, horizontal lines intersecting boxes show the median, and gray points are the mean. Letters indicate significantly different groups of data.

We found no significant differences between vegetative reference point types in terms of mean organic matter pool size, carbon pool size, height, or bulk density for either the live vegetation layer or fibric soil horizon extracted in our unburned cores ($0.3504 < p < 0.9558$). In the unburned areas, the fibric soil horizon ($17.9 \pm 0.63$ cm) was nearly three times thicker than the vegetation layer ($6.19 \pm 0.60$ cm) to the depth that our cores reached (~30 cm; Table 1), which is likely an underestimate of the soil organic layer depth, given that our cores only occasionally reached the mineral soil layer. However, there was larger variation in the vegetation layer ($1\sigma = 6.9$ cm) than fibric horizon ($1\sigma = 2.6$ cm) height. In most cases, the fire burned through the vegetation layer, but never completely through the fibric horizon (Fig. 3, Table 1). The fibric layer (to 24.1 cm average) had almost three times

420  greater organic matter and carbon pool sizes compared to the vegetation layer because of its higher bulk density and

421  height (Table 1).

422  **Table 1.** Bulk density, height, and carbon and organic matter percent and pool sizes of vegetation and fibric layers
423  measured in unburned soils averaged across reference points.  Carbon and organic matter pool measurements are the
424  product of bulk density and their percent contents and are normalized to the height of the vegetation layer and 10
425  centimeters in the fibric layer.  Sample size is 72, which is equally attributable to the two reference points.  Error is
426  reported as the standard error of the mean (SEM).

| Layer | Bulk Density ($kg\ m^{-3}$) | Height (cm) | Percent Carbon (%) | Carbon Pool ($kg\ m^{-2}$) | Percent Organic Matter (%) | Organic Matter Pool ($kg\ m^{-2}$) |
|---|---|---|---|---|---|---|
| Vegetation | $26.7 \pm 1.4$ | $6.19 \pm 0.60$ | 43.04* | $0.712 \pm 0.037$ | 100* | $1.65 \pm 0.09$ |
| Fibric | $54.6 \pm 5.0$ | $17.9 \pm 0.63$ | $39.41 \pm 0.56$ | $2.07 \pm 0.16$ | $88.12 \pm 1.17$ | $4.48 \pm 0.30$ |

427  *Based on assumption
428
429      Carbon and dry organic matter loss estimates similarly increased with burn severity (Fig. 3, Table 2).

430  Based on a two-factor ANOVA with vegetative reference point nested in burn severity, average organic matter and

431  carbon loss differed between burn severity categories (carbon loss: $p < 0.0001$; organic matter loss: $p < 0.0001$) (Fig.

432  3).  Similar to burn depth distributions, the moderate severity range overlapped with the high severity and low

433  severity/unburned distributions (Fig. 3).

434  **Table 2.** Carbon and organic matter loss averaged within each burn severity category across reference point, and
435  total carbon and organic mass loss over each burn severity category area.  Total loss per burn severity category is
436  summed to calculate total loss over the total fire area by burn severity classification.  Error is reported as SEM,
437  except for total loss over the fire area which is reported as the sum of errors.
438

| Severity | Sample Size | Area (hectares) | Carbon Loss | | Dry Organic Matter Loss | |
|---|---|---|---|---|---|---|
| | | | Average ($kg\ m^{-2}$) | Total (Tg) | Average ($kg\ m^{-2}$) | Total (Tg) |
| Low/unburned | 36 | 13,001 | $1.11 \pm 0.10$ | $0.144 \pm 0.013$ | $2.51 \pm 0.22$ | $0.326 \pm 0.029$ |
| Moderate | 36 | 8,516 | $1.62 \pm 0.10$ | $0.138 \pm 0.009$ | $3.62 \pm 0.22$ | $0.309 \pm 0.019$ |
| High | 38 | 32,637 | $1.93 \pm 0.12$ | $0.629 \pm 0.039$ | $4.29 \pm 0.23$ | $1.40 \pm 0.08$ |
| Total | 158 | 54,154 | | $0.911 \pm 0.039$ (Tg) | | $2.04 \pm 0.09$ (Tg) |

439
440      The 2015 fires burned a total of 54,154 hectares in the YKD.  The distribution of dNBR values for our

441  focal fires in the YKD was intermediate compared to dNBR values from all tundra fires and on the low side but

442  within the range of dNBR values for all boreal fires in Alaska between 1989 and 2019 (Fig. A8).  Roughly 60% of

443  the burned area was classified as high severity, while moderate severity and low severity/unburned accounted for

444  about 16% and 24% of the burned area, respectively (Table 2).  As a result, total organic matter and carbon loss

445  within the high severity burn classification was over three times greater than the other two burn severity categories.

446  Although moderate severity occupied less fire area than the low severity/unburned, it contained roughly equal total

447  organic matter and carbon loss to the low severity/unburned areas due to about 50% higher organic matter and

448  carbon losses per unit area on average.  However, it should be noted that losses from low severity/unburned areas

449  are likely overestimates, given the inclusion of unburned areas within regions of low severity burn.  Summing

450  carbon loss and dry organic matter loss over the area of each burn severity category yielded a total loss of 2.04 ±

451  0.09 Tg of dry organic matter and 0.911 ± 0.039 Tg of carbon.  This magnitude of total loss corresponded to an

452  average 3.76 kg m$^{-2}$ of organic matter and 1.68 kg m$^{-2}$ of carbon loss across the fire area.

453  The age of carbon increased by ~50 years from 5 to 20 cm in depth in the unburned area.  Surface samples

454  from burned soils in a low severity/unburned area were dated to 2009, indicating a burn depth of about 5 cm or less,

455  which is within the range of our low severity/unburned burn depth measurements (Table 3).  The rate of change in

456  carbon age between 0 and 5 cm was 2.4 yr cm$^{-1}$; for the 15 to 20 cm interval, it was 3.6 yr cm$^{-1}$.  Because none of

457  our burn depth measurements exceeded 20 cm, all carbon released from the fires was likely relatively new carbon,

458  younger than 64 years of age.  Our average burn depth of 10.3 cm across burn severity and reference point indicated

459  an average age of about 23 years for carbon released.

460  **Table 3.** The fraction of the modern standard and calibrated calendar age for each radiocarbon sample.  Calendar
461  ages are the intercepts for the fraction modern value with the calibration curve.  We do not consider the oldest
462  calendar ages.

| Burn Status | Depth (cm) | Fraction Modern | Calendar Ages CE |
|---|---|---|---|
| Burned | 0 | 1.0486 ± 0.0020 | 1957, 2009 |
| Burned | 0 | 1.0493 ± 0.0024 | 1957, 2009 |
| Burned | 0 | 1.0403 ± 0.0018 | 1957, 2008, 2009 |
| Unburned | 5 | 1.0625 ± 0.0020 | 1957, 2007, 2008, 2009 |
| Unburned | 10 | 1.1198 ± 0.0023 | 1958, 1996 |
| Unburned | 15 | 1.4603 ± 0.0024 | 1963, 1974 |
| Unburned | 20 | 1.0124 ± 0.0017 | 1956 |

463
464  The radiative forcing for gaseous and aerosol emissions from the 2015 fire season was positive for 80 years

465  post-fire according to our model, regardless of the future atmospheric gas concentration scenario (Fig. 4a).

466  However, the magnitude of the fire emissions' radiative forcings decreased with increasing ambient atmospheric gas

467  concentrations, with RCP 8.5 having the lowest radiative forcing at 80 years post-fire (Fig. 4a).  Given the lack of

468  qualitative difference across future scenarios, below we present results for the RCP 4.5 scenario only (Table 4, Fig.

469  4b).  Nitrogen oxides were the only gas with a negative radiative forcing at the 80-year cumulative mean.  However,

470  they had a positive radiative forcing at the 20-year time horizon (Table 4).  The only other negative radiative forcing

471  came from the indirect effect of aerosols that completely counterbalanced the positive radiative forcing from other

472  gaseous emissions in the first year, but the magnitude of this effect in terms of cumulative integrated forcing

473  diminished rapidly after the first decade (Fig. 4b).  $CO_2$ produced the highest radiative forcing through 80 years

474    (Table 4, Fig. 4b).  Nitrous oxide sustained its positive radiative forcing; however, the radiative forcing was of

475    intermediate magnitude with respect to all gaseous emissions.  In contrast, ozone and carbon monoxide exhibited

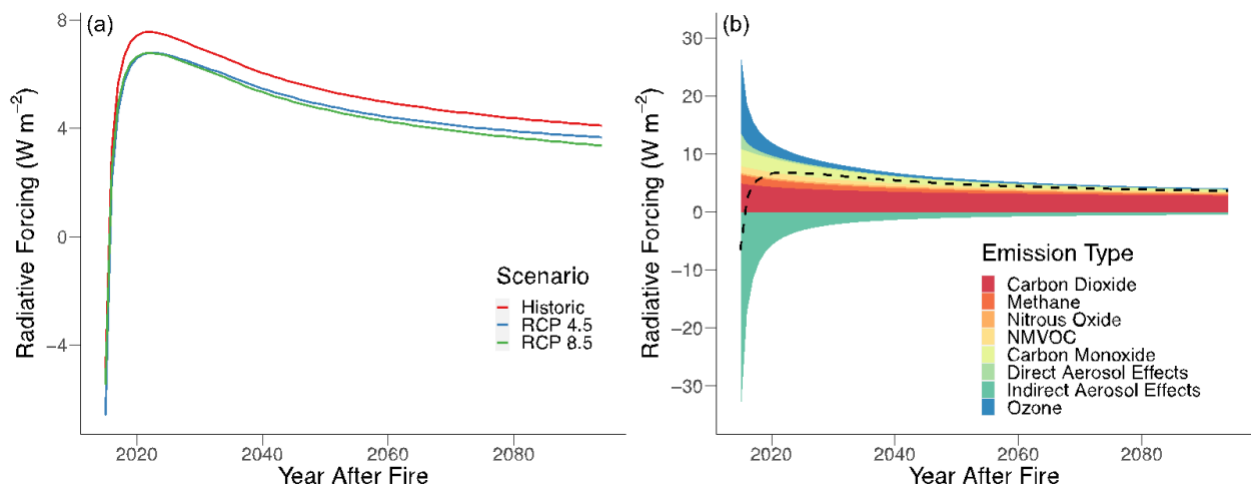
476    high radiative forcings, but the magnitude of these diminished rapidly with respect to carbon dioxide.  Likewise, the

477    effect of the other ozone precursors, aerosols, and methane decreased relatively rapidly (Table 4).  Overall, 20-years

478    post-fire, the effect of ozone precursors increased the total radiative forcing by roughly 11%.  However, this percent

479    increase drops to 5% after 80 years, and the ozone precursors and aerosols show little effect on the percent reduction

480    of the total radiative forcing between 20- and 80-years post fire.  Therefore, the positive radiative forcing is

481    sustained by long-lived gaseous emissions, especially carbon dioxide.

482
483    **Figure 4.** (a) Total cumulative mean radiative forcing of gaseous emissions for three future atmospheric gas
484    concentration scenarios, and (b) cumulative mean radiative forcing of different emissions (dashed line is net forcing)
485    under the RCP 4.5 future scenario over 80 years post-fire from the YKD 2015 fire season.
486
487    **Table 4.** Cumulative mean radiative forcing at 20- and 80- year time horizons post-fire for different gaseous and
488    aerosol emissions from the 2015 fire season under the RCP 4.5 future scenario. Percent reduction refers to the loss in
489    radiative forcing when considering an 80-year time horizon compared to 20 years.

| | Radiative Forcing (W m$^{-2}$) | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Time Horizon | $CO_2$ | $CH_4$ | $N_2O$ | $O_3$ | NMVOC | CO | $NO_x$ | Indirect Aerosol Effects | Direct Aerosol Effects | Total with Ozone Precursors and Aerosols | Total without Ozone Precursors and Aerosols |
| 20 years | 3.69 | 0.76 | 0.28 | 0.67 | 0.62 | 1.51 | 0.03 | -1.71 | 0.14 | 5.99 | 5.41 |
| 80 years | 2.87 | 0.24 | 0.22 | 0.17 | 0.19 | 0.40 | -0.01 | -0.43 | 0.03 | 3.67 | 3.49 |
| Percent Reduction (%) | 22 | 68 | 21 | 75 | 69 | 74 | 133 | 75 | 79 | 39 | 35 |

490
491    **4 Discussion**

492        Here we describe a tundra wildfire season that combusted significant amounts of carbon and had a positive

493    warming effect on the atmosphere due to its gaseous and aerosol emissions over an 80-year time horizon.  Our

494    findings suggest that increasingly frequent tundra fire regimes are an under-studied source of global GHG

495    emissions.  The 2015 fire season in the YKD represents a large efflux of carbon from a wetland tundra ecosystem,

496    generally not considered to be highly vulnerable to fires.  In total, we estimate that about 0.911 Tg of carbon was

497    released from 54,154 hectares of burned area in the YKD.  This fire area is roughly three times the 17,000 hectares

498    of tundra area burned on average annually between the years 2001 and 2018 in Alaska, but it only accounts for a

499    little more than half of the 83,000 hectares of Alaskan tundra that burned in the year 2015 (Scholten et al., 2021).

500    Over the same time frame, the 2015 fire season was the largest to burn in the YKD, with the majority of years

501    burning less than 10,000 hectares (Scholten et al., 2021).  Total carbon loss was driven by a similar per unit area

502    carbon emission rate as the 2007 Anaktuvuk River fire (Mack et al., 2011), which was roughly twice the size of the

503    fire season reported here and burned predominately moderate to high severity in a tussock tundra ecosystem (Jones

504    et al., 2009).  While the current status of tundra ecosystems as a sink or source of carbon remains uncertain, the 2015

505    YKD fire season alone released about a tenth of the carbon sequestered annually from tundra sink estimates

506    (Virkkala et al., 2021).  If tundra ecosystems become a source of carbon to the atmosphere under warming,

507    emissions from tundra wildfires may further exacerbate this positive feedback to warming due to a changing balance

508    between respiration and productivity (Natali et al., 2019, Belshe et al., 2013).  Note that we did not account for post-

509    fire effects on net ecosystem exchange of $CO_2$, which tend to be sources to the atmosphere with greater fire severity

510    due to continued soil respiration until sufficient vegetation recovers (Rocha and Shaver 2011a).

511        The tundra fires in the YKD and at Anaktuvuk River both lost roughly 1.7 to 2.0 kg m$^{-2}$ of carbon (Mack et

512    al., 2011).  Carbon loss per area from these tundra wildfires is within the range of total above- and belowground

513    carbon loss from boreal wildfires, approximately 0.5 to 4 kg m$^{-2}$ (Walker et al., 2020a; Walker et al., 2018b; Rogers

514    et al., 2014), and close in value to the median wildfire loss of carbon, 2.54 kg m$^{-2}$, across all land cover types in

515    Alaska between 2001 and 2012 (Veraverbeke et al., 2015b).  Despite the additional combustion of overstory

516    vegetation in forests, carbon loss in both tundra and boreal systems comes primarily from belowground organic

517    matter (Walker et al., 2020a; Walker et al., 2018b; Rogers et al., 2014).  Compared to remote sensing-derived

518    estimates of combustion, our per unit area *in situ* measurements are on the low end but comparable to large-scale

519    means across Alaska between 2001 and 2018 (Fig. A9).  However, more representative field measurements in tundra

520    fires are needed to validate remotely sensed combustion measurements, as the currently available gridded products

521    (Potter et al. 2022; Scholten et al. 2021; Veraverbeke et al. 2017a; Veraverbeke et al., 2015a) are driven entirely by

522 measurements in boreal forests. Although data from the Anaktuvuk River and YKD fires shows carbon loss per unit

523 area can be similar between boreal and tundra wildfires, information on carbon emissions from tundra wildfires is

524 substantially more limited than boreal forest studies (He et al., 2021; Veraverbeke et al., 2021).

525    Given a fire return interval in the YKD of over around 200 years (Sae-Lim, 2019), fires there likely do not

526 burn through all the carbon accumulated since the last fire. However, tundra fires may initiate post-fire successional

527 trajectories, such as increased shrub cover, that may promote shorter fire return intervals (Hu et al., 2015; Rocha et

528 al., 2012). Shorter fire return intervals, in turn, gradually deplete the available soil carbon stocks. For example, He

529 et al. (2021) showed current tundra areas with frequent fire have relatively shallow soil organic layers. Moreover,

530 our radiocarbon data show that as fires burn deeper into the tundra, they release older carbon more rapidly because

531 the density of carbon by age increases. Coupling these burn-prone successional trajectories with projected increases

532 in tundra wildfire occurrence over the next century due to more frequent hot and dry conditions (Hu et al., 2015;

533 Rocha et al., 2012; Joly et al., 2012; Hu et al., 2010), as well as increases in lightning (Chen et al., 2021; Bieniek et

534 al., 2020; Veraverbeke et al., 2017b), tundra wildfires may begin to burn carbon of an older age than the historic fire

535 return interval. In this case, some tundra regions may transition into fire-driven carbon sources, similar to what has

536 been observed in North American boreal forests (Walker et al., 2019).

537    Immediate combustion is not the only impact wildfires have on tundra carbon stocks, nor do measures of

538 carbon loss fully describe the effect of wildfire emissions on the climate. Fires drive higher soil temperatures and

539 deeper active layer depths that can persist over multiple decades after fire (He et al., 2021; Hu et al., 2015; Rocha et

540 al., 2012), which is caused by the loss of insulating vegetation and upper soil layers from combustion (Rocha and

541 Shaver, 2011b). Higher soil temperatures and deeper active layer depths may catalyze permafrost degradation and

542 thermokarst development (Jones et al., 2015) that accelerates the effects of climate warming on tundra ecosystems.

543 Tundra fires also emit multiple gaseous and aerosol species, which have a net positive radiative forcing for at least

544 80 years post-fire. Hence, even if all of the carbon lost from fire in a tundra ecosystem were to be sequestered again

545 over time through increased photosynthesis, the long-term radiative impact of gaseous and aerosol emissions would

546 still generate atmospheric warming and thus positive climate feedbacks, which are amplified by the post-fire

547 permafrost thaw and degradation not accounted for in this study.

548    This positive radiative forcing is driven by sustained radiative forcings from long-lived GHGs, as well as

549 significant contributions from short-lived climate forcers (SCLFs) such as ozone, NMVOCs, and CO. Carbon

dioxide has the highest radiative forcing at both short and long time horizons because it is emitted in the highest

quantity (Akagi et al., 2011) and its concentration decreases relatively slowly over time, especially in the RCP 4.5

and 8.5 scenarios (Joos et al., 2013). Aerosols and $NO_x$ are SCLFs with both positive and negative radiative

forcings due to distinct mechanisms for affecting the Earth's radiative balance. Aerosols exert a direct radiative

forcing by reflecting or absorbing sunlight in the atmosphere and when deposited on high-albedo surfaces, mainly

snow and ice (Bond et al., 2011). Although black carbon aerosols have a positive direct radiative forcing effect and

organic carbon aerosols have a negative direct radiative forcing effect, the magnitude of black carbon's radiative

forcing far exceeds that from organic carbon (Myhre et al., 2013) (note these are combined in our representation).

Indirectly, aerosols serve as the nuclei for cloud formation, which increases the Earth's albedo and exerts a negative

radiative forcing (Ward et al., 2012). Nitrogen oxides have a near-term positive radiative forcing because they act in

concert with NMVOCs and CO to increase ozone concentrations, yet they have a long-term negative radiative

forcing because they decrease the lifetime of $CH_4$ (Collins et al., 2013).

We include a total radiative forcing with and without aerosols and ozone precursors, CO, NMVOCs, and

$NO_x$. The radiative forcing of these ozone precursors and aerosols is uncertain within current literature, especially

when originating in the Arctic, because of their short atmospheric lifetimes and dependence on regional transport

patterns (Bond et al., 2011, Quinn et al., 2008). However, the long-term cumulative radiative effect of these gaseous

emissions is relatively minimal in comparison to the long-lived GHGs and ozone, and both estimates of total

radiative forcing are of the same order of magnitude as previous estimates of the radiative forcing of gaseous

emissions from boreal forest fires (Huang et al., 2016; O'Halloran et al., 2012).

We found that as the predicted concentration of future emissions increases across RCP scenarios

(Meinshausen et al., 2011), the net radiative forcing of gaseous and aerosol emissions from the fires decreases

slightly. Although counterintuitive, this result is due to the dependency of the radiative forcing equations for long-

lived GHGs, namely $CO_2$, $CH_4$, and $N_2O$, on the ambient concentration of a gas. In more extreme future climate

scenarios, the atmosphere generally has higher concentrations of these GHGs, meaning the additional quantity of gas

emitted by fire has a more diluted effect. This effect is largely counter balanced by the fact that land and ocean $CO_2$

sink capacities are diminished in more extreme future climate scenarios, leading to longer atmospheric $CO_2$ lifetimes

(Fig. A10). The net result is that tundra wildfires have a relatively similar positive radiative forcing under all future

emissions scenarios.

578    While the positive radiative forcing due to gaseous emissions we calculate for the 2015 YKD fire season is

579    within the range of previously calculated radiative forcings for gaseous emissions from boreal wildfires (e.g., Huang

580    et al., 2016, Randerson et al., 2006), boreal forests also have significant multi decadal post-fire albedo changes due

581    predominantly to the combustion of overstory trees that expose snow in the fall, winter, and spring (Lyons et al.,

582    2008).  Randerson et al. (2006) and O'Halloran et al. (2012) show that the negative radiative forcing of these albedo

583    changes can exceed the positive radiative forcing of gaseous emissions and generate a net cooling effect for boreal

584    forest fires over long time horizons in specific locations. It should be noted, however, that (i) the net cooling or

585    warming is heavily dependent on combustion levels, pre-fire canopy composition, and time horizon; (ii) fire-induced

586    albedo forcings predominantly impact the regional climate (Rogers et al., 2013), as opposed to globally-mixed

587    GHGs; and (iii) this cooling impact is expected to diminish with future climate change due to decreases in spring

588    snow cover (Potter et al., 2020).

589    In contrast, tundra ecosystems lack an overstory, and hence fires generate relatively short-lived decreases in

590    albedo (Rocha et al., 2012, French et al., 2016).  These albedo changes are only caused by char and the exposed soil

591    surface in the summer because tundra vegetation rarely exists above the snow layer (Mack et al., 2011).  As a result,

592    within four years after the Anaktuvuk River fire, low post-fire surface albedo recovered to its pre-fire reflectance as

593    the vegetation grew over the charred soil surface (French et al., 2016; Rocha and Shaver, 2011b).  The decreased

594    surface reflectance contributes to increases in soil temperature and thaw depth after fire (Rocha and Shaver 2011b).

595    Therefore, the positive radiative forcing caused by tundra wildfire gaseous and aerosol emissions is not offset by

596    post-fire albedo changes as seen in boreal forest fires.  From 2001 to 2018 about 0.5 Mha of boreal forest burned on

597    average yearly compared to a yearly average of 0.017 Mha of tundra in Alaska (Scholten et al., 2021).  Across the

598    entire Arctic region, about 9.0 Mha of boreal forest and 0.66 Mha of tundra burned on average yearly between the

599    years 1997 and 2016 (van der Werf et al., 2017; biome extents defined by Dinerstein et al., 2017).  Given potentially

600    comparable per area emissions from boreal and tundra fires, boreal fires still release more carbon globally.  Future

601    work could compare the total radiative forcing of tundra and boreal wildfires across their global domains, by

602    integrating gaseous, albedo, and other climate forcing effects for each biome.  The predicted increase in frequency

603    of tundra wildfires, their lack of cooling from post-fire albedo, and their contribution to increased permafrost thaw

604    could drive a net radiative forcing higher than boreal wildfires, which in some cases may cool the climate (Oris et

605    al., 2014; O'Halloran et al., 2012; Randerson et al., 2006).

606     For the purpose of calculating the radiative forcing of wildfires in future research, we make our

607     computational code available on Github (see Code and Data Availability section).  Our workflow is applicable to

608     wildfires in other arctic-boreal ecosystems, and it could be applied to any ecosystem globally with the adjustment of

609     emissions factors for the specific material burning (Akagi et al., 2011) and global warming potentials for SCLFs

610     based on the geographic region of the wildfire (Myhre et al., 2013).  As a result, our method of calculating the

611     radiative forcing of wildfires can be used for studies across multiple regions with the appropriate parameters, which

612     can advance a more complete understanding of the effects of wildfires on the climate.

613     Future work could clarify some of the uncertainties that were not explicitly analyzed in our final estimates

614     of carbon loss and radiative forcing.  We did not account for combustion of aboveground tall woody vegetation in

615     our estimates of carbon loss from the YKD fires.  Although woody biomass is relatively sparse on the YKD

616     landscape, it represents an additional source of carbon from combustion (for boreal forest fires, see Walker et al.,

617     2020c).  Despite the lack of difference we found between soil and vegetation matter characteristics across vegetative

618     reference point groups, there could have been variation in these characteristics outside of the primarily peat plateau

619     areas that we sampled.  Burn depth estimated between *Dicranum* reference points was more variable (although not

620     significantly different) than burn depth estimated between *Sphagnum* reference points. This variation could indicate

621     more variable burn dynamics around *Dicranum* or better efficacy of *Dicranum* for measuring extreme burn depths,

622     which highlights the potential bias introduced by choosing specific combustion markers.  Likewise, we eliminated

623     tussock measurements from our final analysis because they were relatively scarce on the landscape and yielded

624     inconsistent results, perhaps due to burn dynamics around tussocks that remain obscure using our combustion

625     measurement method.  Minimizing potential bias by identifying combustion markers and soil characteristics for

626     different land cover types at a higher resolution within a fire, similar to how different combustion markers are used

627     regionally (Walker et al., 2020c; Mack et al., 2011; Boby et al., 2010) would further improve estimates of carbon

628     loss following wildfire in tundra systems.

629     Our assignment of low, moderate, and high severity was based on our particular collection of field

630     measurements and therefore may not translate well to other fire complexes. While dNBR is widely used to assess

631     fire severity, its accuracy in high latitude tundra systems may be diminished by environmental and methodological

632     factors, such as the timing of image selection and rapid post-fire greening (Chen et al., 2020a; Loboda et al., 2013).

633     Other remotely sensed products have shown promise in tundra (e.g., Chen et al., 2020b), but the applicability of

634   these results to other tundra systems remains to be explored. Within low severity areas, there were likely patches of

635   unburned tundra, potentially leading us to overestimate organic matter and carbon loss measurements. We also used

636   a relatively simple method for extrapolating emissions based on fire severity. More involved techniques such as

637   relating mechanism-based geospatial predictors to field plots using statistical techniques and machine learning,

638   which has been employed for boreal forests in Alaska and Canada (Dieleman et al., 2020; Walker et al., 2018b;

639   Veraverbeke et al., 2015b; Rogers et al., 2014), would likely yield more robust estimates for regional emissions. In

640   this context, we strongly recommend increasing the number of tundra fire carbon emission observations in different

641   regions to facilitate synthesis and biome-wide modeling, as has been done for boreal forests (Walker et al., 2020c;

642   Walker et al., 2020b; Walker et al., 2020a).

643         Our radiative forcings model uses data and algorithms from multiple previous studies, including for

644   emissions factors (Akagi et al., 2011), RCP scenarios (Meinshausen et al,. 2011), GWPs (Myhre et al., 2013), ozone

645   and CO concentration ratios (Jaffe and Widger, 2012), and aerosol direct and indirect effects (Ward et al., 2012), all

646   of which may not be geographically, biologically, or temporally specific to the YKD fires. These factors are also

647   known to be variable depending on fuel types, fire severity and lifetime, and atmospheric dynamics at the time of

648   and following fire (e.g., Wiggins et al., 2016; Jaffe and Widger, 2012; Chen et al., 2009). Since emissions factors

649   for tundra burning were not available, we used emissions factors for burning boreal vegetation. Using boreal

650   emissions factors may overestimate emissions from woody vegetation in tundra and thereby increase the mass of

651   certain gasses released. Similarly, boreal emissions factors might also misrepresent other tundra specific gaseous

652   emissions due to different soil and herbaceous vegetation compositions in the two biomes. Future research could

653   integrate the range of variation associated with emission factors to quantify the uncertainty they introduce into

654   radiative forcing estimates and directly measure the emissions factors of biomass combustion in tundra ecosystems

655   for more accurate radiative forcing estimates. We include estimates of radiative forcing with and without SCLFs,

656   because the radiative forcings of SCLFs emitted from the Arctic are uncertain in current literature. Additional

657   studies assessing how regional atmospheric patterns, emission location, and abbreviated atmospheric lifetimes

658   govern the radiative forcing of SCLFs would constrain this uncertainty. Furthermore, our model does not include

659   the potential mitigating or exacerbating effects of post-fire ecosystem changes on gas concentrations, such as

660   changes in post-fire vegetation (Frost et al., 2020; Zhou et al., 2019; Barrett et al., 2012), increased soil respiration

661   (Rocha and Shaver, 2011a), and increased active layer depths and permafrost degradation (Holloway et al., 2020;

662      Jafarov et al., 2013; Rocha et al., 2012).  With reparameterizations specific to individual fires, our model could be

663      made more accurate for future work.

664      **5 Conclusions**

665      To our knowledge, our study is the first to account for both carbon loss and radiative forcing from tundra

666      wildfires.  We develop a method that pairs *in situ* measurements of combustion levels from local pre-fire reference

667      points with remotely sensed burn severity data to scale organic matter and carbon loss to the entire fire area.  Our

668      estimate of the radiative forcing of gaseous and aerosol emissions from the fire event suggests that tundra wildfires

669      positively reinforce climate warming; however, the role of post-fire vegetative regeneration in mitigating these

670      carbon emissions remains unknown. Our results stress the importance of considering tundra wildfires in assessing

671      climate feedbacks and the need for future research that more explicitly discerns the warming effect of fires across

672      the tundra biome.

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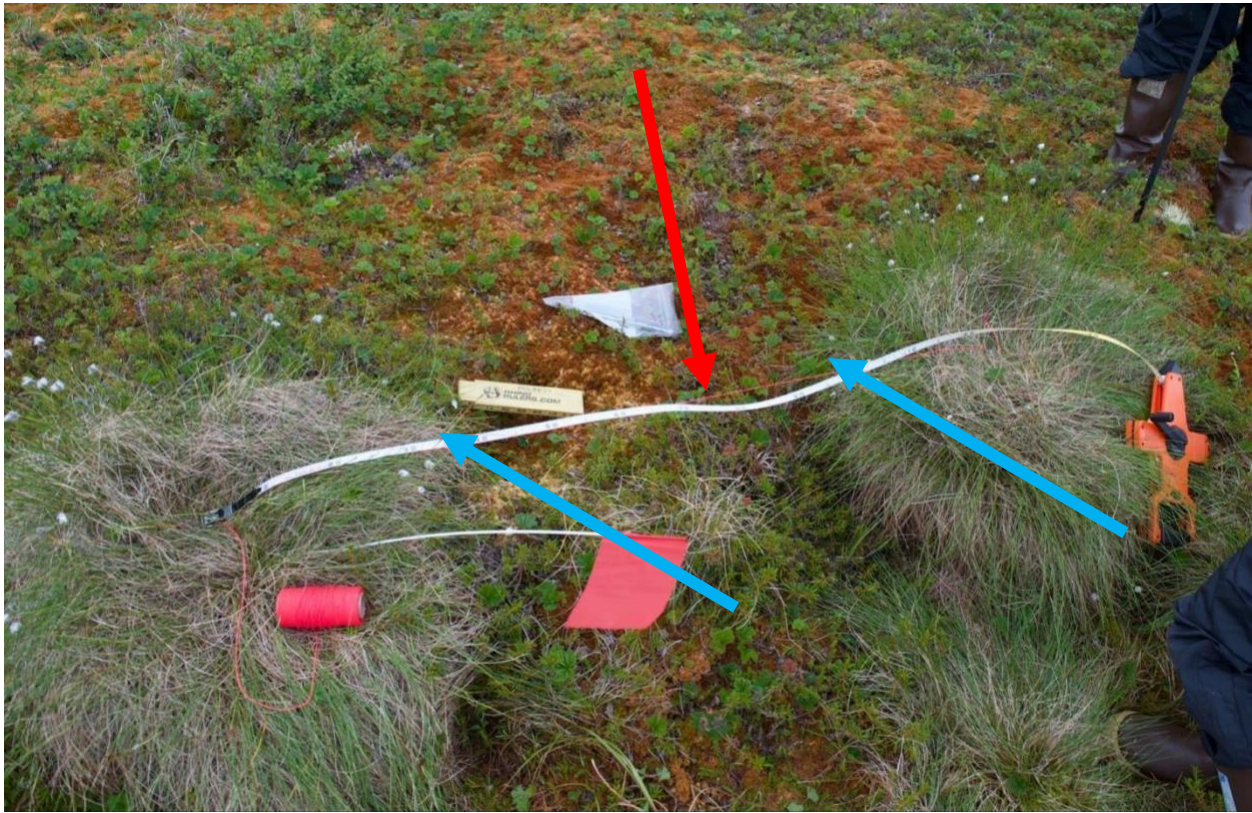
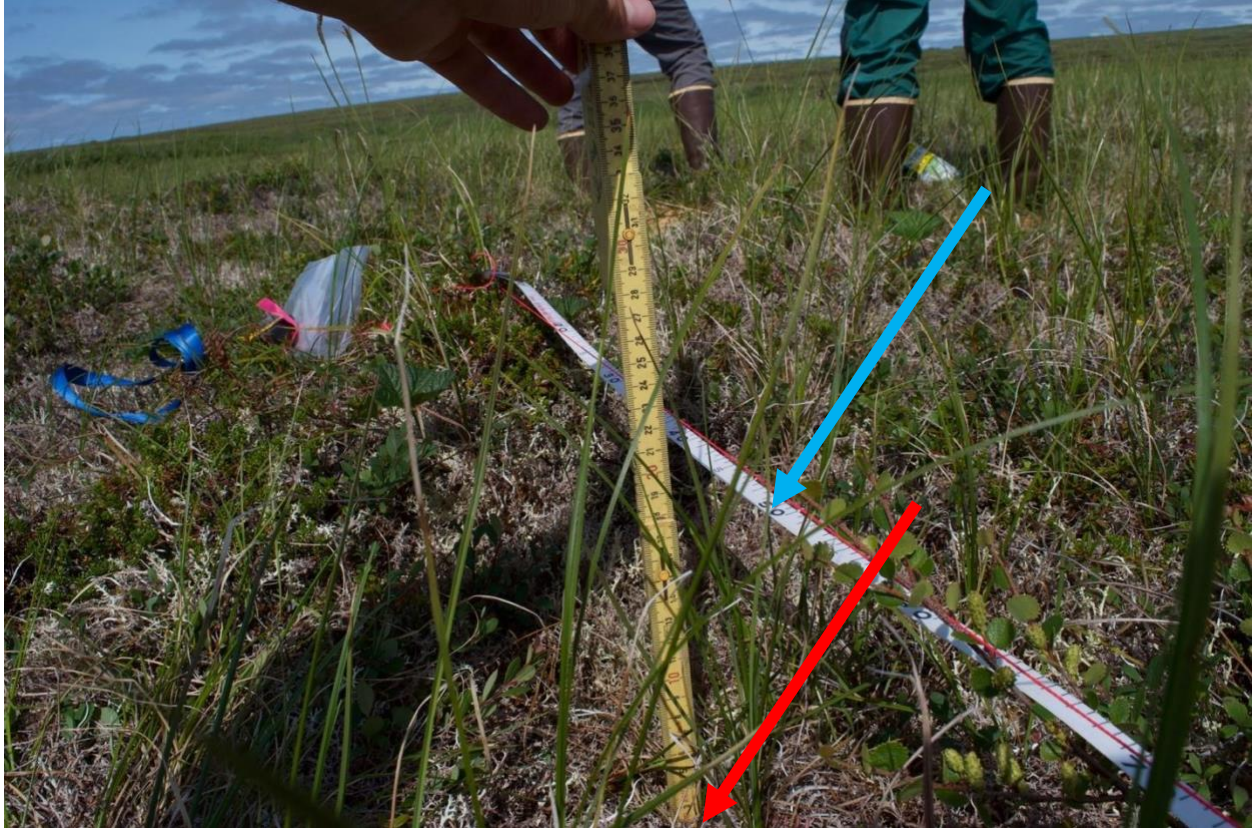
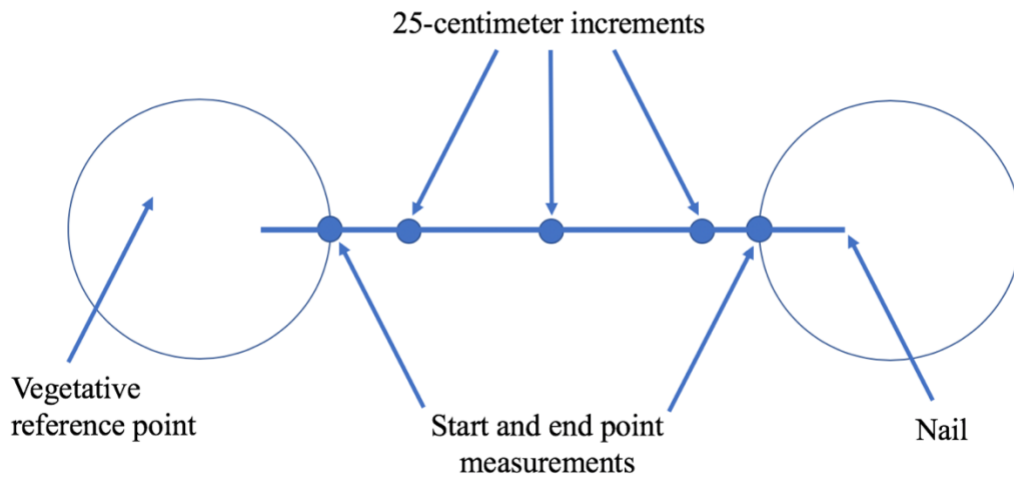
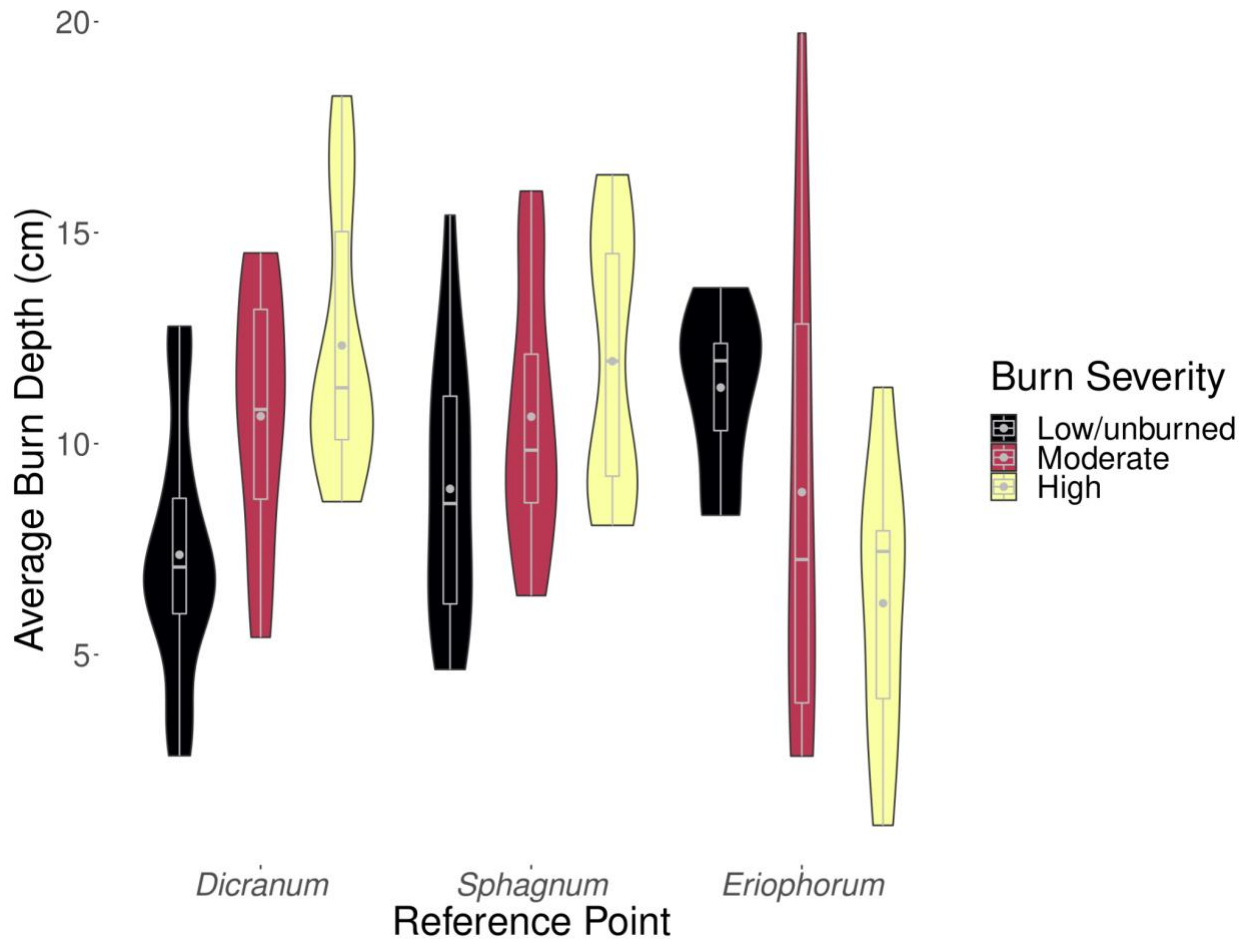
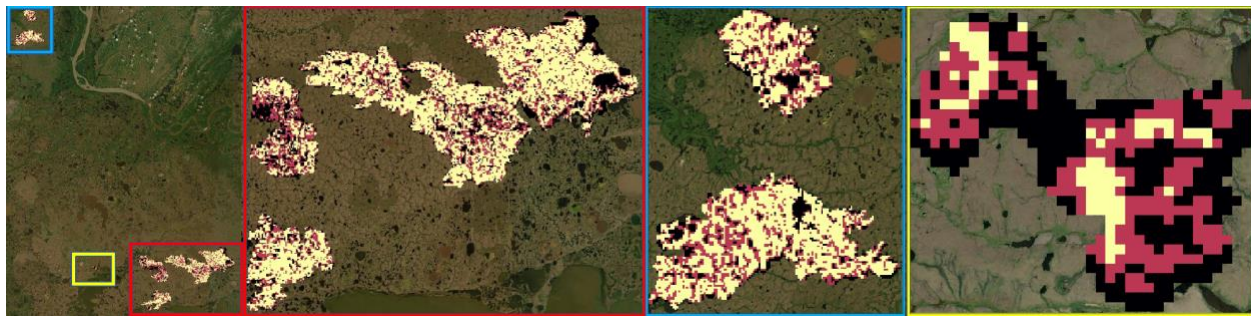
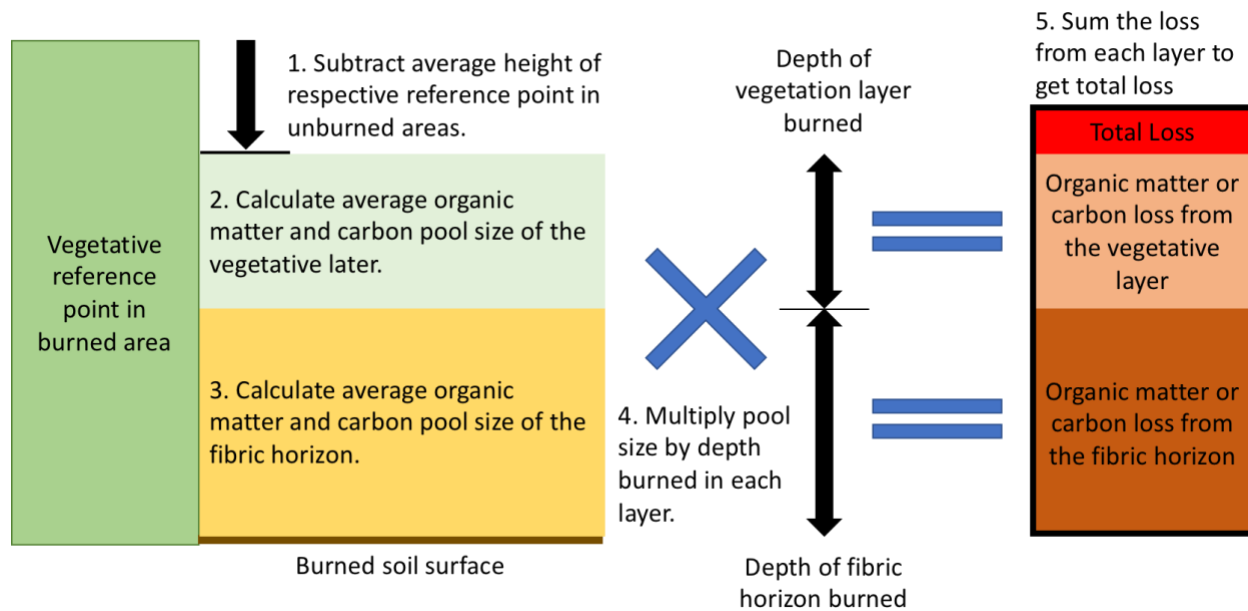
690 **Appendix A**

691
692 **Figure A1.** Height measurement of *Eriophorum vaginatum* in an unburned area from the taut string (red arrow) to
693 the dense vegetation layer between the starting and ending points (blue arrows) at the edge of each tussock patch.
694

695
696     **Figure A2.** Measurement from the taut string to the dense vegetative layer (red arrow) in an unburned area every 25
697     centimeters (blue arrow).

698
699     **Figure A3.** A schematic of the height measurements made along a transect between two vegetative reference points.
700

701
702 **Figure A4.** The distributions of average burn depth per transect within each reference point (including *Eriophorum*
703 *vaginatum*) across burn severity. Boxes encompass the middle 50% of data, whiskers are the upper and lower
704 quartiles, horizontal lines intersecting boxes show the median, and gray points are the mean.
705

706

707 **Figure A5.** Fire severity across the YKD 2015 fire area categorized using dNBR values into low severity/unburned
708 (black), moderate severity (red), and high severity (yellow) categories. The panels show the total fire area (far left),
709 then close ups of the southeastern (red), northwestern (blue), and south-central (yellow) fire scars, respectively.
710 Basemap sources: Esri, DigitalGlobe, GeoEye, i-cubed, USDA FSA, USGS, AEX, Getmapping, Aerogrid, IGN,
711 IGP, swisstopo, and the GIS User Community.

712
713 **Figure A6.** A schematic of organic matter and carbon loss calculations at a single transect in the burned area.
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716

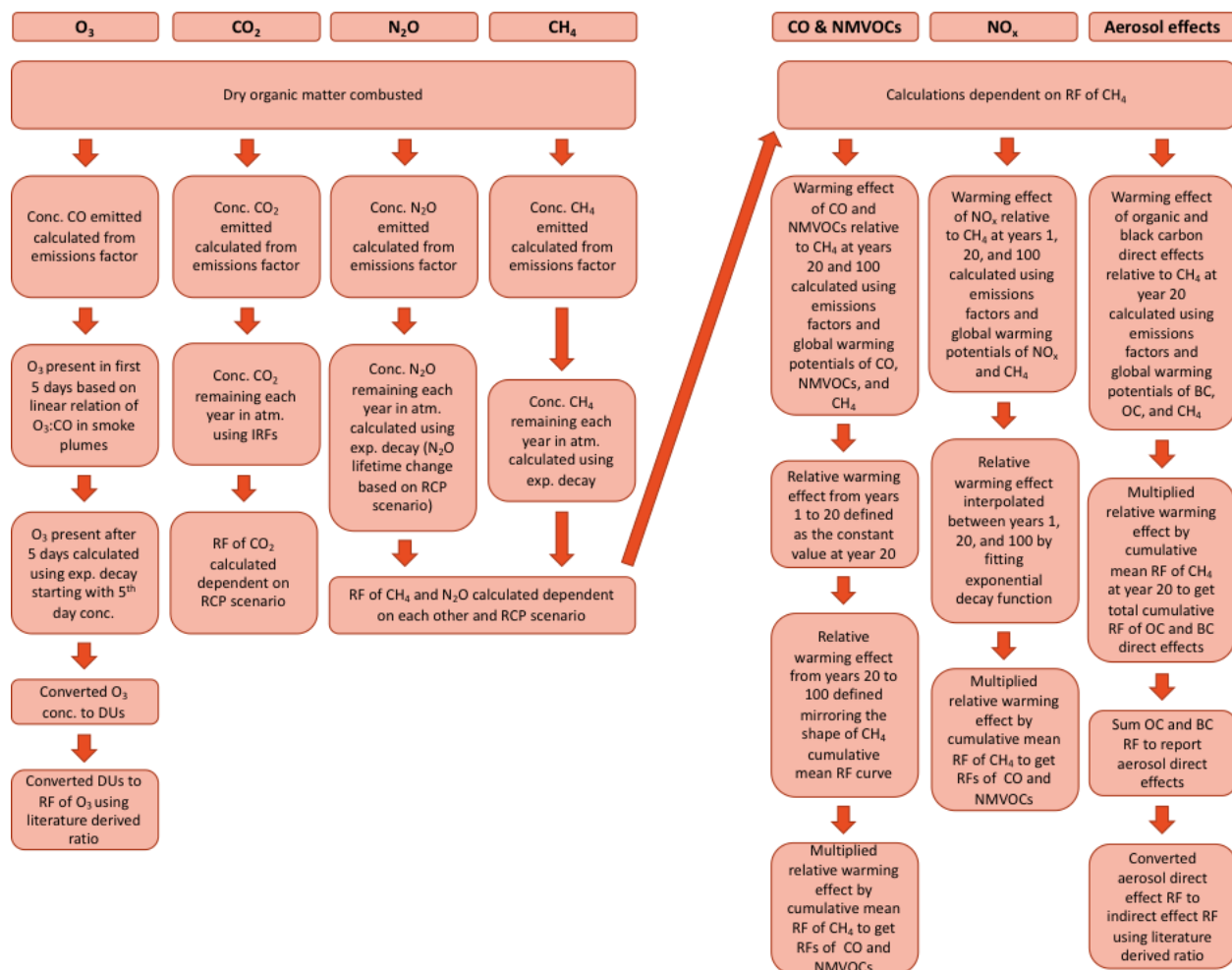



Figure A7. Flowchart of the methodology for the calculation of the radiative forcings of gaseous emissions from biomass combustion generally applicable to any biome with adjustment of biome-specific parameters.

720

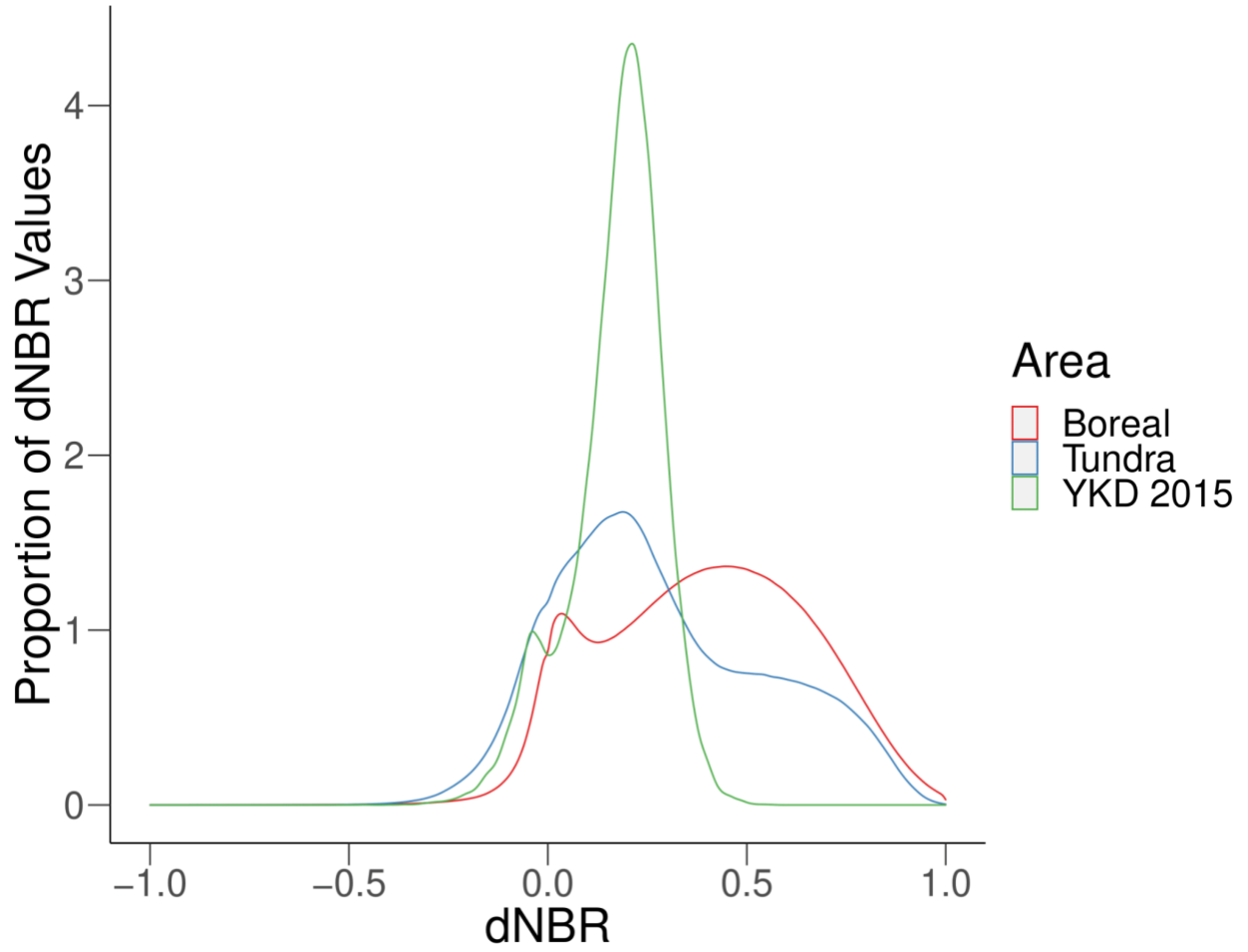
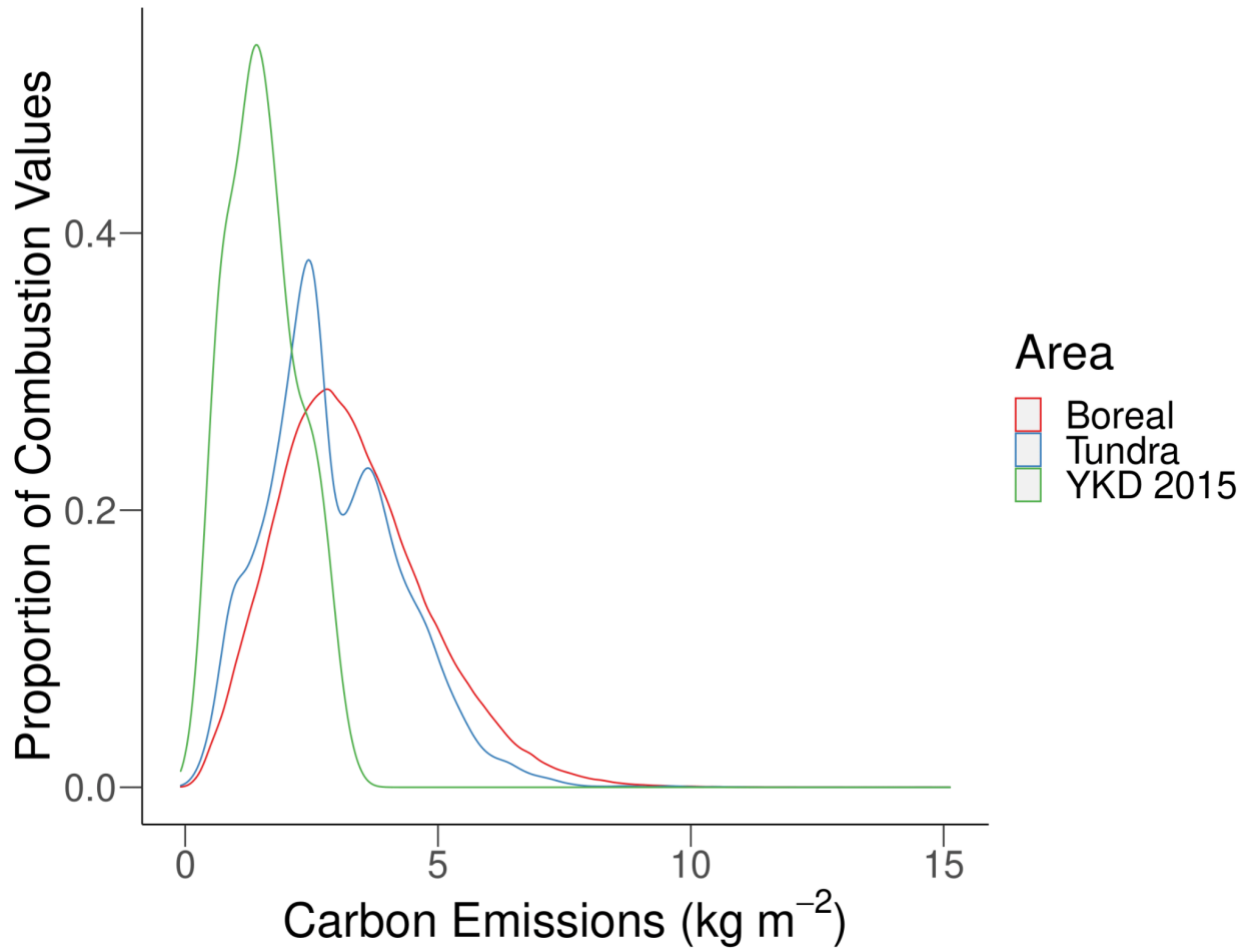



**Figure A8.** The distribution of dNBR values for the 2015 YKD fire season compared to all fires in boreal and tundra systems in Alaska between the years 1989 and 2019.

723
724 **Figure A9.** The distribution of carbon emission values from transect measurements in the YKD 2015 fires reported
725 herein and from remotely sensed pixels of all fires in boreal and tundra systems in Alaska between the years 2001
726 and 2018 (Scholten et al., 2021).

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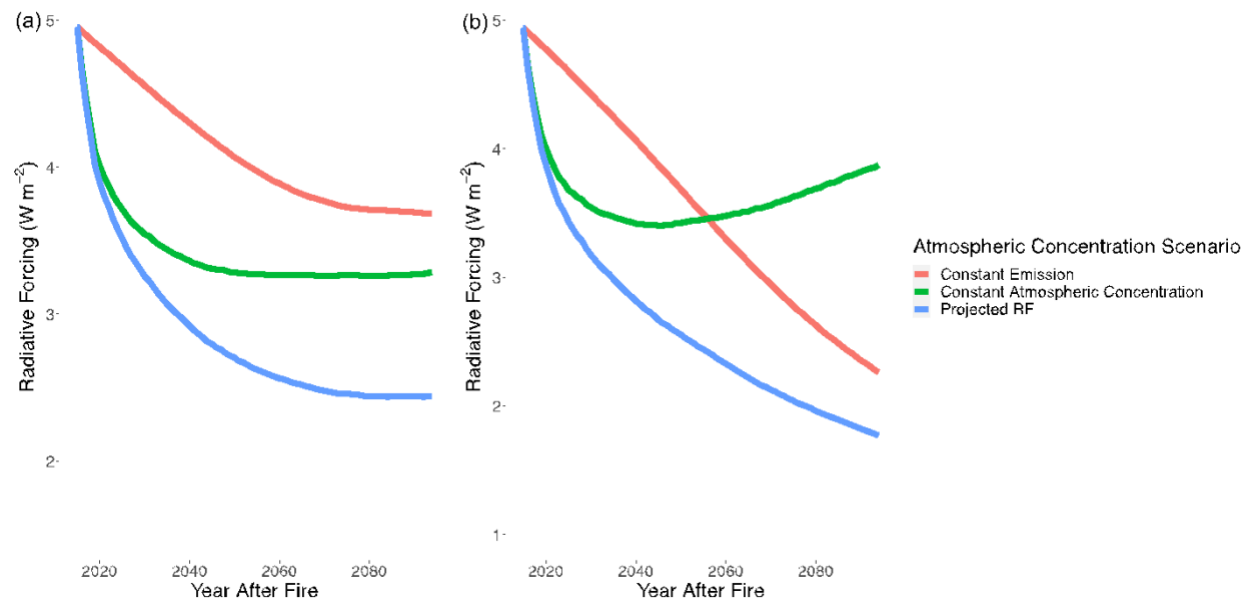
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**Figure A10.** The radiative forcing of $CO_2$ emitted from the 2015 YKD fire season for RCP 4.5 (a) and 8.5 (b) when
(i) holding the emission constant (red lines) so that no emitted $CO_2$ is sequestered, (ii) holding the background
atmospheric concentration constant (green lines), and (iii) the combination of these, resulting in our main $CO_2$
radiative forcing equation (blue lines; Eq. 7). In the case of (i, red lines), increasing atmospheric $CO_2$ concentrations
dampen the radiative impact of fire-emitted $CO_2$. In the case of (ii, green lines), the only time-varying impact on
$CO_2$ radiative forcing is ocean and land sinks, which become less effective and even reverse with more extreme
climate change according to model results presented in Joos et al. (2013). These lines are similar to the impulse
response function of $CO_2$ for their respective scenarios in Joos et al. (2013).

**Code and Data Availability**

The radiative forcings calculation code is available on GitHub and Zenodo via DOI: 10.5281/zenodo.6617455

(Moubarak, 2022). All field and lab raw data used in this study, including vegetative reference point height

measurements and soil and vegetation core metrics, and combustion data products derived from our calculations are

publicly available in the NSF Arctic Data Center via DOI:10.18739/A2PR7MV5P (Moubarak et al., 2020).

**Author Contributions**

MM, BR, SS, and SN were involved with conceptualization, funding acquisition, and methodology. SP completed

investigation of burn severity data and jointly completed visualization with MM. MM, SS, and BR did investigation

of burn depth measurements and soil and vegetation characteristics. MM analyzed data and developed the radiative

forcings model. SS and SN did project administration for the field sampling. MM curated data and wrote the

original manuscript draft. MM, BR, SS, SP, and SN reviewed and edited the final published work.

**Competing Interests**

The authors declare that they have no conflict of interest.

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