LATE HOLOCENE FIRE HISTORY OF THE WATERSHED OF LAKE TANGANYIKA (MAHALE COAST) RECONSTRUCTED FROM MACROCHARCOAL RECORDS

By

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Abstract

Fire in semi humid regions of tropical Africa is driven by random lighting strikes and intentional anthropogenic burnings for charcoal production, which often result uncontrolled fire that causes landscape conversion. However, the relative contributions of localized and strictly human-caused fire vs. more regional burning facilitated by climate events in late Holocene African fire history needs to be clarified. In this study I have investigated the history and drivers of the late Holocene fire, a region in central eastern Africa adjacent to Lake Tanganyika. I have taken a paleolimnological approach, studying sedimented macrocharcoal preserved in sediment cores, to obtain a record of macrocharcoal concentrations and fluxes spanning the last ~ 500 years. I compared macrocharcoal concentrations and flux patterns with lake level and lake temperature changes for nine dated sediment cores (18M, 15M, 20MR, 25M, 21MR, 1A, KH1, 2A, and 6A) collected from nearshore east central coastal areas and southeastern basin locations. From 1500 to 2000 C.E., fires that are recorded in the cores near the northern Mahale area (the east central part of the lake where the majority of records were located) were mostly uncorrelated and presumably represent very localized fires. Only a few events are recorded in multiple cores and thus seem likely to be related to climatic drivers, such as in the late 17th century, a dry period during the Little Ice Age and or during the mid-19th century. Low fire frequencies observed during the late 20th century are possibly related to prior extensive deforestation in the study area. However, the dominance of uncorrelated late Holocene charcoal records near the Mahale area suggests that most of these fires were highly localized and likely human set (for land clearance or charcoal production) rather than being driven by regional climate events.

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1. Introduction

Fire is a critical earth system process that modifies natural environments and biodiversity, affecting human habitats, carbon cycles, and nutrient influxes (Shlisky et al., 2009). It is unclear how many fires occur each year, but human activities contribute to the majority of wildland fires, with lightning strikes also a common trigger of wildland fires (Flannigan et al., 2013). Even though it is difficult to differentiate between human-caused and nature-caused fires, fire has long served as a tool for humans to clear agricultural landscapes and facilitate charcoal production. Frequent human-caused fires can threaten environments and ecosystems, release large quantities of greenhouse gases, and alter landscape hydrology and air quality (Shlisky et al., 2009). These adverse effects of human-caused fire that threaten environments and ecosystems are magnified by rapid human population growth and global climate changes. Therefore, understanding the drivers and distributions of fires are essential pathways to properly managing ecosystems and encountering the challenges of current global climate changes.

Excepting Antarctica, all the continents were settled by humans by the beginning of the Holocene (Marlon et al., 2013). Therefore, studying Holocene fire history worldwide provides us with a means to understand fire-human-climate interactions, and understand the relative roles of climatic and demographic changes in driving fire (Marlon et al., 2013). Globally, fire-dependent and fire-sensitive ecosystems are concentrated in the tropics and subtropics (Nuruddin et al., 2007). Semi humid tropical biomes including deciduous woodlands, savannas, shrublands, and grassland are the most fire-prone regions. Human activities such as landscape conversion to farmland and burning to facilitate charcoal production in tropical regions, especially in Africa, are particularly common causes for uncontrolled fire. In addition to human drivers of extensive fire, climate-driven events, such as extensive aridity associated with the Little Ice Age in eastern

Africa would have increased the likelihood of burning in regions where large fuel loads had previously grown (Palacios et al., 2005). However, the relative contributions of localized and strictly human caused fire vs more regional burning facilitated by climate events in African fire history remains ambiguous.

Fire in nature occurs as a result of interactions between climatic and biological processes that play out over decadal to centurial timescales (Marlon et al., 2013). This makes a paleo-perspective critical for understanding changing fire regimes (Marlon et al., 2013). In this regard sedimentary charcoal records are invaluable for fire reconstructions that span thousands of years and can be interpreted in light of other paleorecords of climate changes and vegetation shifts (Whitlock et al., 2003). Macrocharcoal (here defined as charcoal particles larger than 61µ diameter retained by sediment sieves) as the by-product of fire and biomass burning, in which these larger particles usually result from nearby fires, with delivery through fluvial transport, typically over a short period of time (Graves et al., 2019; Mao et al., 2021). As a result, paleolimnological records of macrocharcoal can provide a high-resolution archive that reflects local fire histories, whether driven by either anthropogenic landscape conversion or and climatic causes.

The focus of my study was to understand fire history from sedimentary charcoal in tropical eastern Africa by studying macrocharcoal deposited in Lake Tanganyika (**Figure 1**), Lake Tanganyika, is located in the eastern side of the African Rift Valley and is surrounded by four riparian countries - Burundi, Tanzania, Zambia, and Congo. Lake Tanganyika is one of the largest (32,600 km^2, volume = 18,880 km^3), oldest (9 to 12 million years), and deepest (Zmax = 1470m, Zmean= 570m) freshwater lakes on Earth (Cohen et al., 2005a). Lake Tanganyika's extensive sedimentary environments provide excellent natural conditions for macrocharcoal deposition, and prior studies have shown that macrocharcoal is abundant in sediment cores collected from the lake (Palacios et al., 2005; Tierney et al., 2010). Fire occurs frequently in the watersheds around the lake, a result of the combination of relatively high rainfall/fuel production coupled with a long dry season. Lake Tanganyika also provides a valuable opportunity to study the potential drivers of fire, whether from human activities or from aridity or lightning strike frequency related to climate over longer time scales.

Wildfire activity around Lake Tanganyika is controlled by various factors. Seasonality of annual precipitation and evaporation rate cycles, coupled with lighting frequency, determines both the duration of wet and drought seasons and the likely timing of ignition, both of which are associated with fire occurrences. Fire and biomass burning normally occurs in the dry season, and vegetation regrowth during the wet season allows renewed buildup of fuel load in the burning area (Krawchuk et al., 2011). The climate around Lake Tanganyika is semi-humid tropical, with annual precipitation ranging between 900 to 1500 mm. The rainy seasons occur between October to April and dry conditions prevail from May to September (Verburg & Hecky, 2003), related to the seasonal migration of the southern tropical rain belt. Vegetation types around Lake Tanganyika also strongly impact patterns of biomass burning. Miombo woodland (dominated by the small-medium-sized deciduous trees *Brachystegia* and *Julbernardia* with grass understory) is the dominant vegetation type in this region of Africa (Frost, 1996). Experiments on African Miombo woodland indicate that this type of vegetation structure results from regular burning of these fire-evolved trees (Furley et al., 2008). Noncultivated low-lands around Lake Tanganyika today are dominated by Miombo woodland, with minor differences in vegetation structure related to precipitation (White,1983; Ivory et al., 2021). In (now rare) undisturbed highlands, around the northern lake shore (the eastern border of Burundi and the northern-west border of Congo), vegetation is dominated by Afromontane forests. On the eastern and southern lake shore (border of Tanzania and Zambia), vegetation is dominated by wetter Miombo woodland. The western shore of Lake Tanganyika (border of Congo) is dominated by wetter Miombo Woodland, a portion of the Miombo Grassland, and Afromontane Forest (Ivory et al., 2021).

The seasonality and the coexistence of Miombo grassland and Miombo woodland might reflect prior biomass burning around Lake Tanganyika. However, it would be inaccurate to correlate biomass burning solely with climate seasonality. The other important factors, human activities such as landscape clearances for agricultural purposes and to facilitate charcoal production (the most important household fuel in the region) are also prominent factors affecting biomass burning around Lake Tanganyika. In central eastern Africa, large-scale land clearance burning was first introduced by humans during the Iron Age mainly for iron production (around 2500 years ago) (Schmidt, 1997). More recently, rapid population growth and human displacement because of regional wars resulted in extensive deforestation and land clearance of populated lowlands (Hall et al., 2009). Those documented extensive land clearances were intensified by the introduction of cassava since 1700. This has leds to large-scale deforestation, especially in the areas of northeastern lake shore where population growth occurred in the 19th C away from slaving activities and the ivory trade (Koponen 1988; Cohen et al., 2005a). Thus, both climate history and human-induced fire histories associated with land use change have to be considered in interpreting the causes of macrocharcoal deposition in my study.

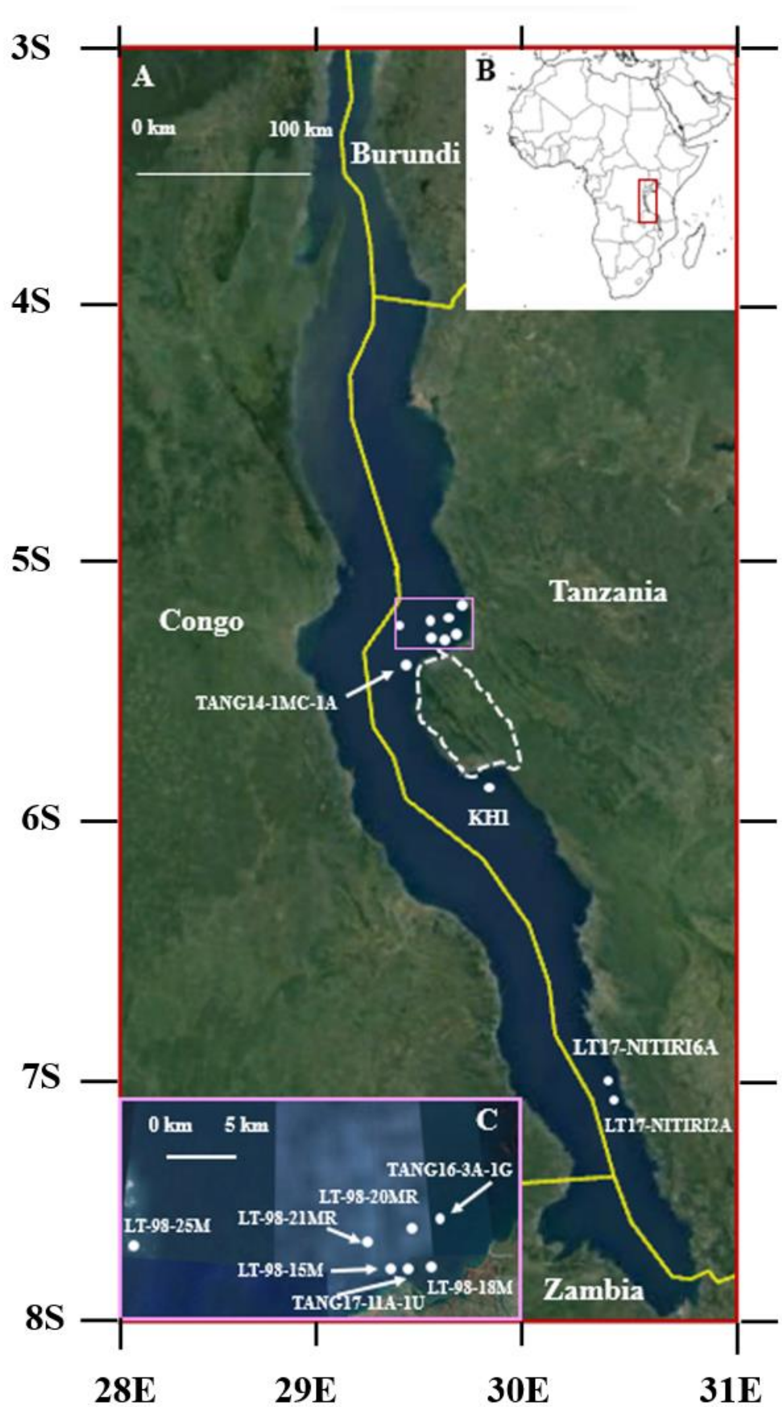

Figure 1 – Panel **A**: Location map of Lake Tanganyika study area and core locations, which includes the yellow national boundaries lines and white enclosed Mahale Mountains Area. Panel **B**: Lake Tanganyika's location in central east Africa enclosed in the red box. Panel **C**: Detailed inset map of the northern Mahale coring sites shown in the inset map. Base maps are from Google Earth.

1.1 - Study Area

For this study, I focused my investigations on sediment cores from the area adjacent to the Mahale Mountains National Park, which is located on the eastern side of Lake Tanganyika (between ~ 6° and 6°30’ S, and 29°40’ to 30°10’E) (**Figures 1 & 2**). The coastal portion of Lake Tanganyika adjacent to the Mahale Mountains area has been subject to extensive past sediment coring activities to understand impacts of deforestation on this landscape (Cohen et al., 2005a; Palacios et al., 2005; McGlue et al., 2021). The Mahale Mountains National Park itself covers 1613 km^2 and has been protected since the early-1960 and established as a national park in 1980 (Cohen et al., 2005a). The forested Mahale Mountain ecosystem is globally important as a protected area for critically endangered chimpanzees (Hunink et al., 2015). The Mahale region outside of the National Park supports a rapidly growing population of local farmers and fishers that live in populated lowland, where the vegetation today consists of miombo woodland, bamboo, and mixed cassava and maize agriculture lands (Hunink et al., 2015; Busch et al., 2018). Due to its diverse natural resources, the Mahale region has been characterized as the Earth’s largest freshwater protected area (Vadeboncoeur et al., 2013). However, the Mahale ecosystem and its biodiversity are under threat due to landscape conversion and deforestation (often using intentionally set fires) caused by human activities. Massive and unrestricted fires lead to landscape conversion from forested woodland to grazing and agricultural lands (Cohen et al., 1993; Patterson, 2000). In addition, the burning of grasslands is done in this region to maintain grass production and pest control, which also leads to increased deforestation (Patterson, 2000). Deforestation has in turn increased soil erosion and sedimentation rates in Lake Tanganyika, threatening aquatic organisms (Cohen et al., 1993). Therefore, the 2013 Tuungane project was introduced to the Mahale region in order to conserve and in some cases restore a healthier

Mahale ecosystem (Vadeboncoeur et al., 2013; Tuungane Project | The Nature Conservancy https://www.nature.org/en-us/about-us/where-we-work/africa/stories-in-africa/tuungane-project/). Therefore, documenting fire and deforestation history and disentangling human from climatic drivers of wildfires is crucial for the future protection and management of Lake Tanganyika's watershed, in particular around the Mahale Mountains National Park.

Figure 2 – Miombo woodlands of the Mahale Mountains coastline, with an overview of the Mahale Coast (Mahale Mountain National Park – andbeyond).

In this study I used macrocharcoal analysis to reconstruct late Holocene fire history around Lake Tanganyika. By applying new and previously collected radiocarbon age data to records of sedimented macrocharcoal collected in cores from Lake Tanganyika adjacent to the Mahale Mountains, I have reconstructed times of increased or decreased fire frequency in the Mahale watersheds to address two questions: (1) What are the possible drivers of high and low fire frequencies among each lake sediment core; and (2) Was fire frequency correlated among the various core sites, as reflected by macrocharcoal fluxes and concentrations in lake sediment cores? Correlated records would be expected if fire frequency was driven primarily by climate variability, which is likely regional in nature. Conversely, fire records that are uncorrelated among watersheds within the regional are more likely to have been the result of localized and probably anthropogenic burning patterns throughout the region. This study provides opportunities to disentangle the puzzle of the potential drivers of late Holocene fires around Lake Tanganyika, which would be valuable for local environmental protection and watershed management in the future, especially under the challenges and pressure of current rapid population growth and displacement in Africa, coupled with regional limited economic resilience to weathering the impacts of climate change (Hall et al., 2017).

2. Methods

2.1 - Lake Sediment Cores Selection and Description

I studied macrocharcoal records from a variety of locations in Lake Tanganyika, positioned at variable distances from shore and in areas with variable amounts of human landscape disturbance in their watersheds. Six sediment cores come from the Mahale region of Tanzania, on the central east coast of Lake Tanganyika (**Table 1, Figures 1 & 3**). Five of these were multicores collected during a Lake Tanganyika Biodiversity Project (http://www.ltbp.org/)-sponsored coring campaign in 1998 (see Cohen et al. (2005a) for further details). Core 1A was collected in 2014 and analyzed in McGlue et al., 2021. Core 6A and 2A was collected in April 2017 (McGlue et al., 2020) and analyzed in Kamulali et al., 2021. Core KH1 was collected by the Nyanza Project Research Experience for Undergraduates program in 2004 and published in Tierney et al., 2010.

Figure 3 – The multicoring device described in Cohen et al., 2005. Panel A shows the multicoring device implementation, panel B shows a core top (sediment/water interface) from a collected sediment multicore.

These multicores were extruded in the field at the time of collection with samples collected every cm and stored in sealed plastic bags in the University of Arizona Paleolimnology Laboratory Cold Room at 4 °C. Sediment cores LT-98-18M, 15M, 20MR, and 21MR were collected from the northeast Mahale Mountain region near the Kabesi watershed. Core LT-98-25M was collected further offshore from the Kabesi watershed. Core TANG14-1MC-1A was collected from Kalilani Bay, where land use is strictly regulated by Tanzania National Parks

(McGlue et al., 2021; Itoh et al., 2007) as a comparison site with the non-managed Kabesi watershed. Additional cores included in this study are from other region in the lake. Core KH1 was collected near the southern part of Mahale Mountain region, in a region that is less influenced by human activities (Tierney et al., 2010). Core LT-NITIRI6A and 2A were collected in the southern basin of Lake Tanganyika offshore from the Tanzanian coastline (McGlue et al., 2021; Kamulali et al., 2021) (**Table 1**).

Along the northeast Mahale Mountains shoreline, from north to south, rivers Rukoma, Lagosa, and Katumbi flow into Lake Tanganyika (Busch et al., 2018). The Kabesi River delta (core site 20MR, 15M, 21MR, 25MR, 18M) at the time of coring had been partially cleared Miombo woodland converted to banana, cassava, and oil palm cultivated land (Cohen et al., 2005a). Further southward, the Lubulungu watershed is located in the western Mahale Mountain region, mainly covered by lowland forest, bamboo bushland (Cohen et al., 2005a) and much less impacted by recent human activity than the coastal sites north of the Mahale Mountains (Cohen et al., 2005a). Adjacent to the Lubulungu watershed, Kalilani Bay (core site 1A) is also located in western Mahale Mountain. The watershed adjacent to Kalilani Bay is mainly covered by native vegetation away from areas of dense human populations (McGlue et al., 2021). Core site KH1 is located adjacent to the southern Mahale Mountains, where the nearby watersheds are also less influenced by human activities (Tierney et al., 2010). The coastline in the southern Lake Tanganyika basin near core sites NITIRI 6A and 2A contains small drainages derived from higher elevations of the lake basin (Kamulali et al., 2021).

Core Number	Geographical Coordinates	Date Collected	Water Depth	Distance Offshore	Length of the Core	References
LT-98-20MR	5.9610° S, 29.8141° E	1998	105m	3.8km	55cm	This study
TANG16-2A-1G	5.9214° S, 29.8543° E	2016	92m	3.6km	None	McGlue et al., 2021
LT-98-25MR	5.9591° S, 29.6117° E	1998	394m	19.3km	52cm	Rosli et al., 2020
LT-98-21MR	5.9673° S, 29.7750° E	1998	141m	3.4km	40cm	This study
LT-98-15M	5.9736° S, 29.7958° E	1998	106m	1.8km	52cm	This study
TANG17-11A-1U	5.9756° S, 29.8001° E	2017	98m	1.7km	None	McGlue et al., 2021
LT-98-18M	5.9760° S, 29.8169° E	1998	75m	2.4km	42cm	Palacios et al., 2005
TANG14-1MC-1A	6.0071° S, 29.7414° E	2014	95m	1.2km	31cm	McGlue et al., 2021
KH1	6.55° S, 29.9666° E	2004	309m	21.8km	149cm	Tierney et al., 2010
LT-NITIRI2A	7.8961° S, 30.7672° E	2017	420m	5.6km	44cm	Kamulali et al., 2021
LT-NITIRI6A	7.963° S, 30.7359° E	2017	680m	4.3km	67cm	Kamulali et al., 2021

Table 1 – Site information for Lake Tanganyika sediment cores investigated in this study.

2.2 - Lake Sediment Cores Sampling

Macrocharcoal samples were prepared by wet-sieving previously weighed samples at 61µ to remove fines. A separate wet aliquot was weighed, dried at 60 °C, and reweighed to determine water content.

Macrocharcoal particles were counted on a gridded counting dish from the sieved residues at the University of Arizona Paleolimnology Laboratory using a Leica M165C Stereo Microscope (**Figures 4 & 5**). Each sample was counted at least two times to insure replicability.

Figure 4 – Using the Leica M165C Stereo Microscope for macrocharcoal counting.

Figure 5 – Macrocharcoal fragments in microscopic, reflected light view.

2.3 - C^{14} and Other Core Geochronology and Age Modeling

When large macrocharcoal pieces (generally length longer than 2.5mm) were observed, I removed these and placed into a glass vial for C^{14} dating Graphite conversion and radiocarbon analyses were done at the University of Arizona Accelerator Mass Spectrometry Laboratory and Northern Arizona University Accelerator Mass Spectrometry Laboratory. C^{14} and Pb^{210} age data constraints for additional cores that were either counted for macrocharcoal by me or for which prior information was used in my analysis can be found in McGlue et al., 2021 for core

TANG14-1MC-1A; Tierney et al., 2010 for core KH1; Palacios et al., 2005 for core 18M; Rosli et al., 2020 for core LT-98-25M; Kamulali et al., 2021 for core LT-17NITIRI6A and LT-17NITIRI2A.

For each core I either established a new age/depth model relationship (core 21MR) (**Figure 6 & Table 2**), relied on a previously published age model for the core being analyzed (core 18, 25M, 1A, KH1, 2A, and 6A), or in two cases where newly collected radiocarbon data was inconclusive for establishing a new age model, applied the age model from a nearby core (core 15M and 20MR). These age models were then used to calculate nominal ages of each macrocharcoal sample.

Sample Number	Affiliated Lab	Sample ID	Dated Material	d13C value	Fraction Modern	Uncalibrated C^{14} Age (+- 1 σ)	Calendar Age Range (68%)
AA114937	UA AMS	LT-98-15M 27-28cm	Macrocharcoal	-27.9	0.7480+-0.0033	328+-35	397calBC to 206calBC
AA114938	UA AMS	LT-98-15M 49-50cm	Macrocharcoal	-25	0.896+-0.011	1068+-99	1048calAD to 1280calAD
AA114939	UA AMS	LT-98-20M 47-48	Macrocharcoal	-25	0.960+-0.011	1622+-92	1463calAD to 1793calAD
AA115511	UA AMS	LT-98-21MR 5-6cm	Macrocharcoal	-29.6	1.5990+-0.0041	Post – bomb	1963calAD to 1967calAD
2525.1.1	NAU AMS	LT-98-21MR 8-9cm	Macrocharcoal			20927+-65	23406calBC to 23040calBC
AA115512	UA AMS	LT-98-21MR 12-13cm	Macrocharcoal	-25.9	0.3481+-0.0013	8477+-28	7576calBC to 7495calBC
2526.1.1	NAU AMS	LT-98-21MR 12-13cm	Macrocharcoal			21683+-67	24070calBC to 23865calBC
2527.1.1	NAU AMS	LT-98-21MR 13-14cm	Macrocharcoal			27861+-132	30096calBC to 29425calBC
2528.1.1	NAU AMS	LT-98-21MR 17-18cm	Macrocharcoal			1074+- 127	771calAD to 1229calAD
2529.1.1	NAU AMS	LT-98-21MR 21-22cm	Macrocharcoal			319+-210	1387calAD
2530.1.1	NAU AMS	LT-98-21MR 26-27cm	Macrocharcoal			56+-91	1796calAD
2531.1.1	NAU AMS	LT-98-21MR 30-31cm	Macrocharcoal			932+-72	1024calAD to 1270calAD
AA115513	UA AMS	LT-98-21MR 38-39cm	Macrocharcoal	-10.9	0.9665+-0.0047	273+-38	1632calAD to 1799calAD

Table 2 – New C^{14} analytical data generated for this project. Dates shaded in blue are those accepted for use in the age model for core LT-98-21MR (blue dots **Figure 6**). Dates shaded in orange are samples yielding plausible but almost certainly reworked charcoal material, that were rejected for use in the age model (orange dots in **Figure 6**). Unshaded dates for core LT-98-21MR are spurious and likely contaminated and not included in further analysis. C^{14} age reports from Northern Arizona University Accelerator Mass Spectrometry Laboratory (NAU AMS) do not include C^{13} correction values and fraction modern values in their reporting.

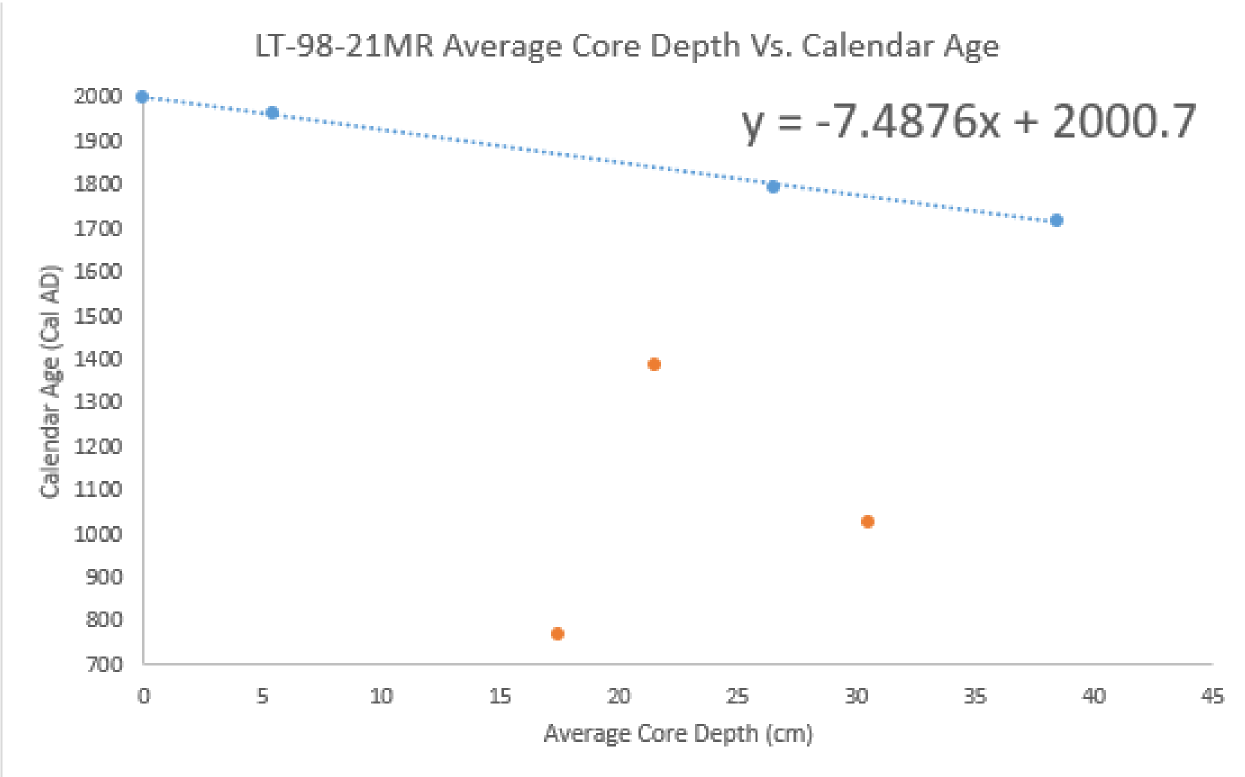

Figure 6 – Age versus depth model for core LT-98-21MR. The blue dots are those accepted for use in the age model for core LT-98-21MR. The orange dots are samples yielding plausible but almost certainly reworked charcoal material, that were rejected for use in the age model for LT-98-21MR.

C^{14} dates for core LT-98-15M and LT-98-20MR were inconclusive and imprecise and are almost certainly derived from reworked old charcoal fragments. Therefore, I applied linear sediment accumulation rates for these cores based on rate determinations from nearby, previously studied cores. For core LT-98-15M, I used the age model from the nearby TANG17-11A-1U "Nkonkwa" core (Distance between core site = 0.9 km mean linear accumulation rate = 0.28cm y^{-1}, McGlue et al., 2021). For LT-98-20MR, I applied rate data from the TANG16-2A-1 G "Mgambo" core (Distance between cores sites = 2.6 km, mean linear accumulation rate = 0.15cm y^{-1}, McGlue et al., 2021).

2.4 - Charcoal Concentration/Charcoal Flux/Calendar Age Calculations

I estimated both macrocharcoal concentration (numbers of particles gm^{-1} dry weight) and macrocharcoal flux (numbers of particles $cm^{-2}yr^{-1}$) for samples from each sediment core to determine possible fire frequencies on each core site. Macrocharcoal concentration (numbers of macrocharcoal/gram of dry weight) indicates how many microcharcoal particles occur in every gram of dry weight sieve samples and calculated by the following formulae:

Macrocharcoal Concentration = Number of macrocharcoal particles ÷ Dry Mass (1)

Number of Macrocharcoal fragments = Mean of all macrocharcoal counts on the same sample (2)

Dry Mass Sieved Samples = (Mass of wet sample prior to sieving × (100 – water content loss of sample from drying, expressed as %)) ÷ 100 (3)

Water content loss % = 100 – ((mass of dried sample ÷ sample mass before drying) × 100) (4)

Macrocharcoal flux (macrocharcoal concentration/cm^2/year) indicates the amount of macrocharcoal deposited over a given area and during a given time period. Macrocharcoal flux is a more meaningful estimator of fire activity than simple concentration because of the interacting effect of background sedimentation rate dilution of charcoal concentration. Macrocharcoal flux is calculated by the following formulae:

Macrocharcoal flux = Number of microcharcoal particles cm^{-2} $year^{-1}$ (5)

Number of macrocharcoal particles $cm^{-2}year^{-1}$ = Number of macrocharcoal particles $gram^{-1}$ × Mass (gram) of sediment deposited $cm^{-2}year^{-1}$ (6)

Number of microcharcoal particles $cm^{-2}year^{-1}$ = Macrocharcoal particle mean count ÷ Sample dry weight (7)

Mass (gram) of sediment deposited $cm^{-2}year^{-1}$ *= Volume of sediment deposited* $cm^{-2}year^{-1}$ *(8)*

Volume of sediment deposited $cm^{-2}year^{-1}$ *= Volumetric percent sediment × Linear sedimentation accumulation rate* ($cm^{-2}year^{-1}$) *(9)*

Volumetric percent sediment = Proportion sediment ÷ (Proportion sediment + Proportion water) (10)

Proportion sediment = Sediment mass/gram deposition (11)

Proportion water = Water content ÷ 100 (12)

Sediment deposition = (1- (Water content ÷ 100)) (13)

Proportion water = Water content loss % ÷ 100 (14)

2.5 - Data resampling/Correlation Analysis

I separately graphed macrocharcoal concentrations and fluxes versus time (Calendar Age C.E.) for each sediment core (20MR, 15M, 21MR, and TANG14-1MC-1A) and compared them with previously studied charcoal records from cores LT-98-18M (Palacios et al., 2005), LT-98-25M (Rosli et al., 2020), KH1(Tierney et al., 2010), Nitiri 2A and 6A (Kamulali et al., 2021) (**Figure 7**). I also compared the charcoal records with previously published lake level (Cohen et al., 1997; Alin et al., 2003) and TEX_{86} based lake water temperature (Tierney et al., 2010) reconstructions. However, because of the different temporal scales of these earlier studies it was necessary to linearly resample each data set to uniform time increments (10 & 20 years) to permit a correlation analysis (**Figure 8**) to determine if fire events were correlated between sites and with potential external drivers. As a proxy for regional precipitation and temperature, warmer temperature is frequently associated with increased precipitation in tropical Africa.

Therefore, comparisons between fire events with lake level and lake temperature are helpful to determine whether fire events were driven by regional climatic events. Correlation analysis was done using the Pearson correlation function in PAST 4.03 (Hammer et al., 2001) and plotted in EXCEL to determine resampled macrocharcoal flux correlations between each pair of core sites.

3. Results

3.1 - Macrocharcoal Concentrations and Fluxes (**Figure 7**)

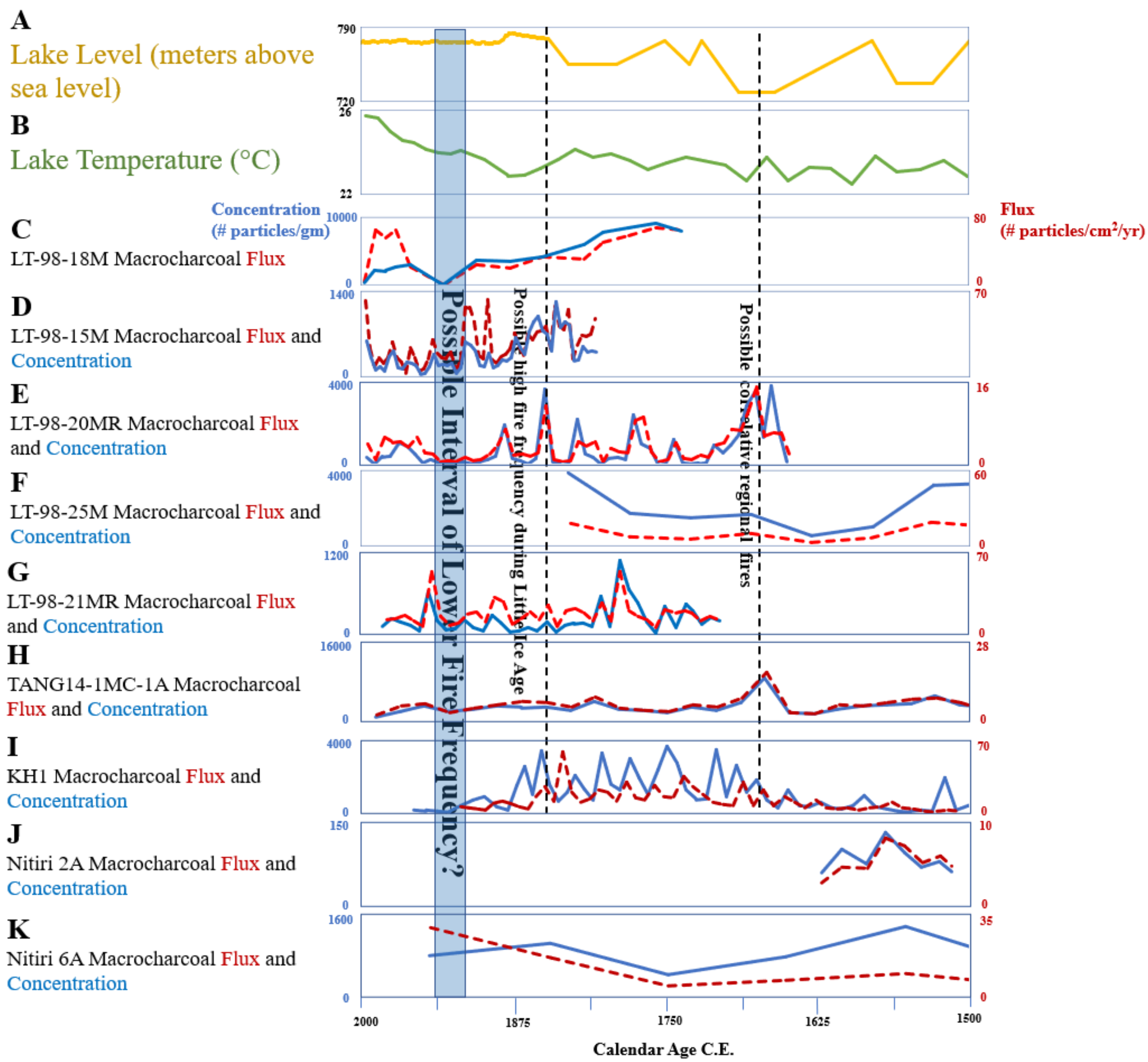

Figure 7 – Comparison of macrocharcoal flux and concentration from the study cores with lake level (Cohen et al., 1997 & Alin et al., 2003) and lake temperature (Tierney et al., 2010) for the past 500 years. Macrocharcoal concentration indicated by blue lines, and flux indicated by red dashed lines.

LT-98-18M (Palacios et al., 2005)

Maximum values in macrocharcoal flux in Core 18M occurred from ~ 1740 - 1800 C.E. After 1800 C.E., flux continuously decreased to no macrocharcoal in 1931 C.E. After that date, macrocharcoal flux increased to another peak in 1988, after which it decreased up to the end of the core record.

LT-98-15M

Peaks in macrocharcoal flux in Core 15M occurred from 1830 – 1850, 1896, 1910-1914, and 1996 C.E. at the top of the core. Macrocharcoal flux values were generally low throughout the 20th Century except for at the very end of the core record, where high values were again observed.

LT-98-20MR

Peaks in macrocharcoal flux in Core 20MR occurred in approximately 1647, 1770, 1848, and likely 1881 C.E. macrocharcoal flux after 1881 C.E. decreased to nearly no macrocharcoal in ~ 1988 C.E. and increased slightly at the core top. As with Core 15M flux rates were generally low throughout the 20th Century.

LT-98-25M (Rosli et al., 2020)

Macrocharcoal flux changes are relatively muted in Core 25M, with the highest flux rates occurring ~ 1529 and 1826 C.E. The record from this core does not extend beyond the mid-19th Century.

LT-98-21MR

Significant peaks in macrocharcoal flux in Core 21MR occurred in ~ 1787 and 1944 C.E. At other times flux rates were relatively low throughout the core (core record ends in 1983).

TANG14-1MC-1A

A single major macrocharcoal flux peak occurs in this core in ~ 1790 C.E. As with several other cores, microcharcoal flux is quite low throughout the 20th Century.

KH1 (Tierney et al., 2010)

Peaks in macrocharcoal flux in Core KH1 occurred in ~1684, 1734, and 1835 C.E. Macrocharcoal flux decreased after ~1835 C.E. to nearly no macrocharcoal at the core top, again show the “low 20th C charcoal” pattern evident in most other cores.

Nitiri 2A & Nitiri 6A (Kamulali et al., 2021)

Core 2A shows minor macrocharcoal flux alterations from ~ 1515 to 1621 C.E., with the highest macrocharcoal flux in the mid-16th Century (~1553 C.E. and 1569 C.E.). Because of a core-top hiatus this record does not extend beyond the early 17th Century. In the nearby Core 6A, a low resolution microcharcoal record shows low flux values in the macrocharcoal flux values in the 16th - 18th Centuries increasing to a highest macrocharcoal flux in ~ 1944 C.E. at the core top.

In general, there is no obvious relationship between mean macrocharcoal flux rates and distance from shore and human disturbance for the adjacent watersheds, which suggested macrocharcoal patterns record localized fire history.

3.2 - Correlation Analysis

The correlation analysis of resampled macrocharcoal flux among each core, lake level and lake temperature in an even time increment (10 & 20 yrs) shows that the core records for the most part are uncorrelated in time both with each other and with either lake level or lake temperature, demonstrating that each area producing charcoal making up the individual core records had a unique fire history which was not fundamentally driven by climatic events (**Figure 8**). The only exceptions to this pattern are in the significant correlations between cores 25M and 18M, 1A and 20MR, KH1 and 15M, and 6A and 18M. However, there was no evidence to indicate those correlations were related to geographic proximity and degree of disturbance.

	LT-98-18M	*LT-98-15M*	*LT-98-20MR*	*LT-98-25M*	*LT-98-21MR*	*1A*	*KH1*	*2A*	*6A*	*Lake Level*	*Lake Temp*
LT-98-18M		0.90777	0.49754	0.0059408	0.58495	0.25985	0.059832		1.98E-06	0.12283	0.94491
LT-98-15M	0.045353		0.66946		0.95061	0.078981	0.016956		0.15549	0.06255	0.17878
LT-98-20MR	0.2069437	-0.104776		0.88314	0.84312	0.0043773	0.88668		0.1266	0.34825	0.094952
LT-98-25M	-0.970874		0.0575243		0.26683	0.1682	0.20748	0.45437	0.060307	0.26347	0.41753
LT-98-21MR	-0.16726	0.0157275	0.0391723	0.5417935		0.38767	0.81781		0.47977	0.14561	0.79086
1A	-0.337217	0.6133627	0.654296	0.3501958	0.250517		0.78892	0.060281	0.37896	0.27324	0.12435
KH1	0.6123842	0.6707746	-0.028213	-0.322024	-0.052123	-0.060558		0.73816	0.52463	0.8781	0.063147
2A				0.3408541		0.7341599	0.1561341		0.10825	0.55457	0.72403
6A	-0.963087	-0.598725	-0.412413	0.4645172	0.2261163	-0.192462	-0.143305	0.6578234		0.60658	0.035299
Lake Level	-0.450011	-0.435221	-0.480507	-0.297291	-0.282253	-0.58401	0.0244001	-0.30656	0.5657339		0.0036129
Lake Temp	0.0213094	-0.322026	-0.286706	0.209247	-0.052476	-0.315446	0.2860468	0.1653233	0.4407662	0.4080059	

Figure 8 – Linear r Pearson correlation analysis of the resampled macrocharcoal flux. The red line separates the P-values (above the red line) and R2 values (below the red line) for each core pair. Yellow shading indicates significant correlation (p<0.05) and blue shading indicates the corresponded R2 values.

In an earlier study, Tierney et al., (2010) found a weak inverse relationship between lake water temperature and charcoal flux over longer time periods than considered in this study. However, this was not observed in this study: comparing macrocharcoal fluxes and macrocharcoal concentrations on each core and correlation coefficients between resampled macrocharcoal fluxes with resampled lake level and lake temperature, showed no obvious relationships between charcoal concentration and flux and lake level or lake temperature changes (**Figure 7**). Due to the general wetter climate in East Africa, lake levels were higher during first half of the Little Ice Age (1250 to 1550 C.E.), (Alin et al., 2003), and between 1500 to 1550 C.E., the lake level was around 770 msl (modern lake level is around 776.4 msl). From 1500 to 1850, lake levels fluctuated but with a generally decreasing trend, reflecting more arid conditions in the lake basin during the second half of the Little Ice Age (Alin et al., 2003). After the Little Ice Age, lake levels started to rise due to rainfall recovery (Alin et al., 2003), and in 1878 C.E., the lake level reached its highstand at 784 msl (Evert, 1980).

Between 1500 to 1900 C.E., Lake Tanganyika's water temperature in the mid-lake Kalya area (record) varied between 22.5 to 24.3 degrees Celsius. By contrast, between ~1900 to 2000 C.E., Lake Tanganyika warmed considerably by nearly 3 degrees Celsius (Tierney et al., 2010) reflecting the effect of anthropogenic greenhouse gases on global climate (Tierney et al., 2010) and consistent with instrumental records for much of this same time period (O'Reilly et al., 2003). However, there is no true statistically consistent trend in macrocharcoal flux at all sites over this same time period that related to lake level and lake temperature changes in this study.

4. Discussion

Generally, fire histories near the Mahale area between 1500 to 2000 C.E., reflected by the macrocharcoal concentration and flux patterns and macrocharcoal flux correlation analysis, suggest localized fire as the more dominant part of the signal. The correlations between climate proxies and fire history are generally weak and statistically insignificant.

The general lack of correlation between individual fire records notwithstanding, most common broad macrocharcoal concentration and flux peaks appeared asynchronously between sites from the late 17th to mid – 19th Centuries. And from the early 16th to mid - 17th Centuries, macrocharcoal flux and concentration values generally appeared low (except for core 25M) compared with the other periods in the northern Mahale areas. There are several time periods that stand out in terms of patterns of increased or decreased fire activity, which may be reflecting regional, climate driven controls. One possible interval of more widespread fire activity in the 1660s – 1670s is evident in cores 20MR, 25M, 1A, and KH1. A second possible regional pattern of higher fire frequency interval occurred ~ 1850 C.E., at the end of the Little Ice Age evidenced in cores 15M, 20MR, and KH1. Both of these events might be related to the drought during the second half of the Little Ice Age (1550 to 1850 C.E.). In addition, much of the 20th Century, and in particular from ~ 1920-1930 represents a possible interval of lower fire frequency as evidenced by macrocharcoal flux data from numerous cores (18M, 15M, 20MR, 21MR, and 1A,). The low 20th Century fluxes of microcharcoal particularly in core top samples suggest that fire activity has not increased in proportion to rising human population density, possibly because of prior extensive deforestation in the study area.

Excepting these few possible correlative fire frequencies, there were no more correlated macrocharcoal concentration and flux patterns observed in my study. Also, combined with the results from the correlation analysis, the resampled macrocharcoal flux from each of the cores are rarely correlated. Therefore, explanations of macrocharcoal concentration and flux patterns besides regional climate also need to be properly considered. Palacios et al., (2005) proposed that the drivers for the local scale fires could be varied, such as intentional human-induced burning for agriculture and grazing purposes, cooking fires, and random uncontrolled wildfires. Since most of the macrocharcoal records from my study do not align with each other, as a more dominant result, those non-correlated marcrocharcoal fluxes and concentrations might be indicative of locally set fires (for land clearance or to prepare wood for charcoal production) instead of regionally extensive and/or simultaneous fires triggered by a combination of large fuel loads followed by drought and regional fire activity. A combination of random lightning strikes that burned small areas of forests, coupled with uncorrelated burning by people for land clearance and small-scale charcoal production offers a plausible explanation for this pattern. In addition, the uncertainties of the age model on each sediment core might affect the distributions of macrocharcoal concentration and flux patterns. Taken as a whole, it is unlikely that the macrocharcoal concentration and flux patterns in my study primarily reflect regional patterns of correlated fire histories in the Lake Tanganyika watersheds driven primarily by climate, and it is more likely that the macrocharcoal concentration and flux patterns reflect individual fires in individual adjacent watersheds that fill into the core records. Individual small-scale fire events appear to have dominated the fire history in the Northern Mahale area during the late Holocene.

5. Summary

This study of macrocharcoal flux patterns from Lake Tanganyika sediment cores reveals a pattern of fire history that can be explained by a combination of dominantly local anthropogenic and secondary regional climatic drivers around Lake Tanganyika in the late Holocene (1500 to 2000 C.E.). The few regionally correlated fire events evidenced in my study notwithstanding most of the fire peaks are uncorrelated, and most likely related to localized land clearance activities and the production of cooking charcoal and probably anthropogenic burning. Most common broad macrocharcoal concentration and flux peaks appeared asynchronously between sites from the late 17th to mid – 19th Centuries. Macrocharcoal flux and concentration values generally appeared low from the early 16th to mid - 17th Centuries. A few correlated regional fire events linking cores 20MR, 1A, and KH 1 in the late 17th century; and cores 15M, 20MR, and KH1 in the mid-19th century might be related to dry periods during the Little Ice Age. Similarly, a general decrease in macrocharcoal flux after the mid-19th century might be related to low macrocharcoal production due to prior extensive deforestation in the Northern Mahale area.

At this point, I cannot explicitly provide the exact drivers for the historical fire events in the Northern Mahale area by observing macrocharcoal concentration and flux patterns and the correlation analysis of resampled macrocharcoal flux. However, the difference in timing of the distributions of macrocharcoal concentration and flux peaks from my study, even for sites that are closely adjacent, indicates that these macrocharcoal histories record predominantly watershed-scale, localized wildfires.

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