NOVEL DISTURBANCE REGIMES IN THE ARCTIC: PALEOECOLOGICAL PERSPECTIVES ON FIRE AND THERMO-EROSION FROM ALASKAN TUNDRA

BY

MELISSA LYNN CHIPMAN

DISSERTATION

Submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy in Ecology, Evolution, and Conservation Biology in the Graduate College of the University of Illinois at Urbana-Champaign, 2017

Urbana, Illinois

Doctoral Committee:

Professor Feng Sheng Hu, Chair, Director of Research
Professor Thomas M. Johnson
Professor Bruce L. Rhoads
Assistant Professor Surangi W. Punyasena
ABSTRACT

Anthropogenic warming is amplified in the northern high latitudes, and disturbance processes such as wildfire and thermo-erosion (i.e., ground subsidence transport of ice-rich sediments) have increased in frequency and magnitude in tundra ecoregions in recent decades. These novel disturbance regimes could increase the role of the Arctic in exacerbating climate warming through the release of large carbon stocks stored in frozen soils. Fires can directly release soil carbon through combustion, and catastrophic thermo-erosion can rapidly increase the carbon pool available for microbial decomposition. On longer timescales, fires can enhance permafrost thaw by reducing surface albedo, enhancing surface roughness, and facilitating shrub expansion and snow drift, which can alter the soil thermal regime. Thus, these novel disturbance regimes may result in more rapid ecosystem changes in the Arctic compared to warming alone. However, the paucity of observational data from remote tundra ecoregions, spatial heterogeneity of ground-ice deposits, and rare burning in many tundra ecoregions at present limits our understanding of the drivers of and potential interactions between fire and permafrost disturbance. I overcome these challenges with paleoecological techniques to examine these disturbance processes over a broad range of scenarios, which is vital given the novel climate and vegetation settings predicted for the future. The main questions of my dissertation research are:

Chapter 2 - How do modern tundra fire regimes compare to long-term natural variability?

Chapter 3 - How does catastrophic permafrost thaw vary through time and what are the dominant drivers? Chapter 4 - What is the relationship between wildfire and thermo-erosion?

Paleorecords of fire and thermo-erosional activity provide critical information regarding natural variability, which is difficult to assess given the rarity of modern tundra burning and the spatial variability of catastrophic permafrost thaw. To address my first question (Chapter 2), I used macroscopic charcoal stored in lake sediments to reconstruct past fire regimes from sites located in several Alaskan tundra ecoregions, which span a variety of modern vegetation and climate combinations. These paleofire records, spanning the past ~10,000 to 35,000 years, are among the first from these remote tundra ecoregions, and show that the spatial pattern of tundra burning on the Alaskan landscape has been in place for millennia, suggesting that ongoing changes to the fire regime will have profound impacts to ecosystem dynamics and carbon storage in historically undisturbed regions. For Chapter 3, I used sophisticated geochemical techniques, including X-ray fluorescence, X-ray diffraction, and strontium isotopes to identify past episodes
of watershed thermo-erosion and examine the drivers of these events over the past 6000 years. This record of thermo-erosion from the Alaskan North Slope provided one of the first reconstructions of thermo-erosional activity from the Arctic. Intervals of shoreline thaw over the past 6000 years broadly corresponded to periods of high summer temperature, illustrating strong climatic controls on thermo-erosion in tundra regions that rarely burn. In addition, the lack of thermo-erosional episodes at a nearby site over the same time period suggests that positive feedbacks facilitate catastrophic thaw in ice-rich areas where it has previously occurred.

These multi-millennial records of fire and permafrost activity provided critical information on the long-term dynamics of Arctic disturbance regimes, as well as a template to assess future change. However, interactions between these processes may further enhance the impacts of Arctic warming. To assess the relationship between fire and thermo-erosion, and gain a deeper understanding of the factors that facilitate catastrophic permafrost thaw, I combined paleoecological and soil analyses from sites in the Noatak River Watershed (NRW), a region that is a potential analogue for Arctic tundra ecoregions in the future (Chapter 4). Using lake sediments from the NRW, I reconstructed past episodes of shoreline thermo-erosion and watershed fires to directly test the relationship between these disturbance regimes using superposed epoch analysis (SEA). The SEA shows a significant relationship between watershed fires and thermo-erosional episodes over the past 3000 years, illustrating a key feedback mechanism that exacerbates catastrophic thaw. Moreover, thermo-erosion occurred several decades after fire events, suggesting long-term feedbacks between fires, landscape-scale vegetation structure, ground-heat flux, and permafrost dynamics. This record provides new insight into how these novel disturbances interact, suggesting that climate-driven changes to Arctic fire regimes can facilitate catastrophic permafrost thaw.
Dedicated
in loving memory to my parents
Frank William Chipman Sr.
and
Lillie Mae Chipman
ACKNOWLEDGEMENTS

It is a daunting task to thank all the amazing people that have supported me in completing my PhD. I owe a huge thank you to the Environmental Protection Agency and the National Science Foundation for funding this research. I also thank the members of my doctoral committee; Dr. Bruce Rhoads, Dr. Tom Johnson, and Dr. Surangi Punyasena, for their endless patience and sound advice during the past six years. Most of all, I would like to thank my advisor, Dr. Feng Sheng Hu, for taking a chance on a first-generation college student with a funny accent. Thank you for always holding me to a higher standard and for pushing me to achieve more than I ever thought possible.

In addition to my committee, I have also been fortunate to receive amazing training and mentorship from others at UIUC and beyond. I would like to give a special and heartfelt thanks to Dr. Carol Augspurger for training me to become an effective teacher, and for showing me what is possible in undergraduate education. Also, a big thank you belongs to Dr. Carolyn Dash, whose professional advice has been hugely helpful to me through the years.

I would especially like to thank Dr. Phillip Higuera at the University of Montana for his past and continuing mentorship. Thank you for teaching me about fire ecology, charcoal analysis, and field work. But most importantly, thank you for taking such beautiful field photos, and for showing me how to tie impressive knots.

I would also like to thank the professors and administrative staff of the Program in Ecology Evolution, and Conservation Biology, the Department of Plant Biology, and the School of Integrative Biology here at the University of Illinois for their ongoing support and dedication. I would especially like to thank Bill Flesher for the awesome chats and encouragement through the years, Jessica Katterhenry for helping me to manage my fellowship, and Rayme Ackerman for always having the right answer to every question.

A huge thank you belongs to the past and present members of the Hu Lab Group, who were my surrogate family during the past thirteen years. You are all wonderful people who taught me so much, and I will remember each and every one of you!

I would especially like to thank Denise Devotta, who was the Frodo to my Samwise. We started this journey together, all the way back in the Shire (which is the office we shared when working on our M.S. degrees). It has been a rare privilege to be your friend and to share this winding road with you.
In addition, I cannot imagine who I would be today without Dr. Ben Clegg, who took me under his wing when I first showed up in the Hu lab in 2004. You taught me how to think like a scientist, always encouraging me to go one step further. My happiest times as a graduate student were sitting with you at my desk, writing papers line by line, and eating tiramisu.

A big thank you to Dr. Ryan Kelly, for not only being an awesome collaborator and scientist, but also for being a great friend. Not many folks would listen so patiently to my rants and consistently laugh at my terrible jokes. Thank you for adopting me into your awesome group of friends, which soothed my occasional bouts of homesickness. And most of all, thanks for singing ridiculous songs with me in the field.

A special thanks to Dr. Mike Urban for being an amazing travelling companion on long road trips, and for being an awesome cook. You will always be my favorite curmudgeon!

Also, thanks to Matias Fernandez for forcing me to occasionally leave my apartment, and to Joe Napier and Dr. Guillaume de Lafontaine for making me take lunch breaks (and for getting me addicted to Bosco sticks).

Thanks also to the amazing undergraduate lab technicians, Cassie Stephens and Richard Vachula, without whom this work would not have been possible. Thank you for all your hard work…and for listening to me go on and on (and on) about permafrost thaw.

Last but not least, I would like to thank my siblings and closest friends for their years of support. Thanks to my sister Chunkie for teaching me to read and write, my brother Willie for never backing down and for always being himself, and my little brother Billy, for being one of my best friends (and for not holding that busted tooth against me). I would also like thank my Aunt Debbie and Uncle David, who made a special point of looking out for me after my parents died. I love you both so much and treasure the cards you send me. Thanks also to my dearest friends, Jesse Yuhasz and Chris Stinson, for sticking with me for the past 20 years, and keeping my spirits up with your hilarious and heartfelt phone calls.

And finally, thanks to Scott Yuhasz for making me smile, writing sweet notes, and cooking dinner. But most of all, thanks for always being on my side.
# TABLE OF CONTENTS

CHAPTER 1: INTRODUCTION ........................................................................................................... 1

CHAPTER 2: SPATIOTEMPORAL PATTERNS OF TUNDRA FIRES: LATE-QUATERNARY CHARCOAL RECORDS FROM ALASKA ........................................... 17

CHAPTER 3: MULTIPLE THERMO-EROSIONAL EPISODES DURING THE PAST SIX MILLENNA: IMPLICATIONS FOR THE RESPONSE OF ARCTIC PERMAFROST TO CLIMATE CHANGE ................................................................. 40

CHAPTER 4: LATE-HOLOCENE INTERACTIONS BETWEEN CLIMATE, FIRE, AND THERMO-EROSION: IMPLICATIONS FOR NOVEL TUNDRA DISTURBANCE REGIMES ...................................................................................... 56

CHAPTER 5: CONCLUDING REMARKS ......................................................................................... 91

APPENDIX A: SUPPLEMENTARY MATERIAL FOR CHAPTER 2 .................................................. 94

APPENDIX B: SUPPLEMENTARY MATERIAL FOR CHAPTER 3 .................................................. 97

APPENDIX C: SUPPLEMENTARY MATERIAL FOR CHAPTER 4 .................................................. 107
CHAPTER 1: INTRODUCTION

*Tundra ecosystems and disturbance regimes*

Land surface temperatures in the northern high latitudes are projected to increase by as much as 8.3 °C by the end of the 21st century (RCP8.5 scenario; Collins et al., 2013). This amplified warming (Miller et al., 2010; Serreze and Barry, 2011) has already resulted in profound environmental changes in the Arctic, including sea-ice reductions (Stroeve et al., 2008; Wang and Overland, 2012), glacier retreat (Arnedt et al., 2002), lengthening of the growing season (Høye et al., 2007), increasing biomass and shrub expansion (Tape et al., 2006; Elmendorf et al., 2012; Epstein, 2012), and widespread permafrost thaw (Jorgenson et al., 2006; Lawrence et al., 2008). The northern circumpolar permafrost zone comprises only 16% of the world’s terrestrial land area, yet contains almost 50% of the world’s soil organic carbon stocks, and nearly twice as much carbon as the modern atmospheric pool (Tarnocai et al., 2009). Thus, these rapid ecosystem changes may exacerbate the release of permafrost carbon to the atmosphere and amplify Arctic warming via positive feedbacks (Chapin et al., 2005; Lenton, 2012). In addition, the Arctic tundra biome, located primarily in the continuous permafrost zone (i.e., >90-100% land area underlain by permafrost; Brown et al., 1997), contains large reserves of soil carbon in yedoma (Zimov et al., 2006) and massive ground ice deposits (Fritz et al., 2015). These tundra permafrost and ground-ice carbon stocks may be highly labile and subject to rapid decomposition when thawed (Vonk et al., 2013; Abbott et al., 2014; Schreiner et al., 2014; Strauss et al., 2015), and are thus particularly vulnerable to ongoing warming.

Climate-driven disturbances have been intensifying throughout the Arctic in recent decades, facilitating broad-scale ecological and geomorphological changes in tundra ecoregions (Rowland et al., 2010). These disturbances include *press disturbances* such as shrub expansion and climate-driven permafrost thaw, and *pulse disturbances* such as rapid permafrost degradation and wildfire (Grosse et al., 2011). Wildfires are a relatively rare phenomenon in tundra ecosystems compared to highly flammable boreal forests because tundra regions are characterized by cooler temperatures, low fuel (biomass), and limited ignitions (lightning) (Wein, 1976; Kaschekche and Turetsky, 2006). Despite this generalization, over 2.2 million ha of Alaskan tundra has burned in the past 60 years, and recent studies suggest that climatic warming and associated vegetation change may dramatically increase future fire occurrence (Hu et al., 2015; Young et al., 2016). Enhanced burning may modify physical properties in the tundra by
removing organic soil cover (Liljedahl et al., 2007; Mack et al., 2011), altering surface radiative energy fluxes and soil thermal properties (Chambers et al., 2005; Rocha et al., 2012), promoting changes in vegetation composition (Racine et al., 2004), and increasing permafrost thaw depths (Rocha and Shaver, 2011; Zhang et al., 2015). Coupled with the direct impacts of climate warming on permafrost soils, these post-fire processes may alter the role of tundra ecosystems in the global carbon balance (McGuire et al., 2009; Mack et al., 2011, Hu et al., 2015).

In circumpolar Arctic landscapes, a combination of cold climate and frozen soils has inhibited post-glacial landscape evolution by constraining hydrological flow patterns (Quinton and Carey, 2008) and limiting sediment supply (Kokelj et al., 2017). However, recent increases in the rate and magnitude of permafrost thaw and associated erosional processes (e.g., Jorgenson et al., 2006; Osterkamp, 2007; Segal et al., 2016) suggest that Arctic climate change may relax these constraints and facilitate massive mobilization of glacial sediments and associated old carbon stocks. Deeper permafrost thaw depths have already resulted in an increase in the proportion of ancient terrestrial carbon mobilized throughout the Arctic (Feng et al., 2013), and thus these climate-driven processes may facilitate carbon losses in excess of that from microbial degradation alone (Grosse et al., 2011; Schuur et al., 2015). Landscape subsidence and associated disturbance processes in response to permafrost degradation are collectively called thermokarst (Jorgenson and Osterkamp, 2005). In ice-rich permafrost areas, thermo-erosion, which is the rapid subsidence and lateral transport of previously frozen soils, is often the dominant thermokarst process associated with permafrost degradation. Thermo-erosional features occur in upland tundra regions and include thermokarst gullies (water tracks that form on gently sloping hillsides), active layer detachment slides (thawing atop supersaturated soils that result in massive movement of material along the detachment plane), and retrogressive thaw slumps (erosional features in steep terrain that degrade laterally as the associated ice wedge thaws). Together, these thermo-erosional disturbances can result in widespread and often dramatic geomorphic, hydrologic, and ecological changes in ice-rich tundra landscapes.

**Alaskan tundra fire regimes**

Modern climate-fire relationships in the Alaskan tundra show that warm, dry summers coincide with an increase in fire frequency (Hu et al., 2015; Young et al., 2016). This strong link between climate and wildfire suggests that ongoing Arctic warming will likely increase the
probability of fire in historically low-fire tundra ecosystems, resulting in novel fire regimes. For example, the 2007 Anaktuvuk River Fire on the Alaskan North Slope, which doubled the total area burned north of 68°N in Alaska since 1950 (Jones et al., 2009; Mack et al., 2011), occurred during a record-setting warm and dry summer with reduced sea-ice, and was unprecedented over the past 5000 years (Hu et al., 2010). Similarly, the highly flammable Noatak River Watershed in northwestern Alaska also experienced record high fire years in recent decades (Higuera et al., 2011a; Rocha et al., 2012), with the largest number of fires of the modern record (the past ~60 years) occurring in the warm summer of 2010. Modern burning within the Noatak ecoregion also exhibits spatial variability that is likely related to climatic variation, with the largest observed fires occurring in the warmer western area of the Noatak watershed, which has thinner snowpack and thus earlier spring thaw compared to the eastern area (Macander et al., 2015). This climate-driven spatial variability in fire activity was also recorded in charcoal-based paleofire reconstructions from the western area of the Noatak watershed, which show that cooler up-valley sites burned less frequently over the past 2000 years than warmer down-valley sites (Higuera et al., 2011a). Together, these observations and paleo-fire reconstructions document the link between climate and fire at multiple spatial scales.

Climate also impacts tundra fire regimes indirectly via changes in biomass flammability and vegetation composition. The modern spatial pattern of burning, defined by the ecoregion fire-rotation period (FRP = average time required to burn an area equal to the size of the ecoregion; Johnson and Gutsell, 1994) reflects spatial heterogeneity in both climatic and vegetation controls on the fire regime between Alaska tundra regions. Warm areas in northwestern Alaska, such as the Noatak and Seward Peninsula, have high shrub- and tussock-dominated vegetation cover and experience frequent burning, with modern FRPs of ~425 years. In contrast, less shrubby tundra settings such as the cold and dry North Slope and the warm and wet Yukon-Kuskokwim Delta rarely burn, with ecoregion FRPs of ~1890 and ~4370 years, respectively. Within these Alaskan tundra ecoregions, area burned between 1950 and 2011 was biased towards shrub and tussock tundra vegetation types compared to moist non-acidic tundra vegetation cover (Rocha et al., 2012). Similarly, paleofire reconstructions from the Noatak River Watershed show a relationship between site-specific vegetation composition and fire frequency over the past 6000 years (Higuera et al., 2011b). Recent climate warming has facilitated the expansion of shrubs in historically herb-dominated tundra settings in Alaska (Stow et al., 2004;
Chapin et al., 2005; Tape et al., 2006; Euskirchen et al., 2009). This change in vegetation structure could potentially enhance fire frequency by increasing the amount and flammability of fuels, as suggested by increasing frequency of tundra fires in northcentral Alaska coincident with the transition from herb-dominated to shrub-dominated tundra ~13,500 years ago (Higuera et al., 2008). However, short observational records (~60 years) and the paucity of paleorecords from tundra ecoregions make it difficult to fully characterize ongoing changes in fire regimes and understand long-term climate-vegetation-fire dynamics.

Fire frequency and/or severity may increase in Alaskan tundra ecoregions in response to a combination of increasing summer temperatures (Young et al., 2016), changing atmospheric patterns that facilitate ignition (Romps et al., 2014), and enhanced biomass and/or flammability of fuels (Goetz et al., 2005; Myers-Smith et al., 2011). Fires can alter the thermal regime of soils by changing albedo and increasing the depth of the permafrost active layer, and continued high occurrence and increased severity of fires could potentially change tundra ecosystems from a sink to a source of carbon, releasing large carbon stores to the atmosphere and potentially enhancing greenhouse warming (McGuire et al., 2009; Mack et al., 2011; Hu et al., 2015). Furthermore, altered fire regimes may also impact vegetation structure and facilitate shrub encroachment (Racine et al., 2004) which can have important feedbacks to climate, fire, and permafrost dynamics. Regrowth of vegetation following tundra fire is often rapid (Mack et al., 2011; Bret-Hart et al., 2013). However, lichens and mosses recover slowly following fire (Racine et al., 2004; Jandt et al., 2008), which can alter energy partitioning in the soil (Rocha and Shaver, 2011) as well as diminish food sources for caribou (Jandt et al., 2008; Joly et al., 2012). Thus, understanding these potential feedbacks in tundra ecosystems is critical to anticipate the ongoing suite of environmental changes in the Arctic, and to modify management practices for the preservation of subsidence resources for Arctic residents.

**Thermo-erosional degradation in the Arctic**

Throughout the circumpolar Arctic, continuous permafrost areas have maintained massive ice deposits and glacial sediment stores associated with past glaciations. Recently deglaciated permafrost terrains are extensive, comprising over 7 million km$^2$ of the circumpolar Arctic (Kokelj et al., 2017). Thermokarst-vulnerable areas, which are associated with landscape properties such as ground-ice content, topography, soil texture, landscape history, and vegetation
composition, are estimated to cover 20% of the northern hemisphere permafrost zone and store approximately half of this region’s soil carbon (Olefeldt et al., 2016). In some tundra ecoregions, the rate and magnitude of thermo-erosional features has increased in recent decades concurrent with warming temperatures and increasing active-layer depth in permafrost soils, underscoring the importance in understanding the impact of these features on Arctic ecosystems. For example, both the number and area of retrogressive thaw slumps has increased by 125% and 160%, respectively, over the past ~50 years on Hershel Island, Canada (Lantuit and Pollard, 2008). Similarly, the areal extent of thaw slumping has increased by 36% over the past ~30 years in the Mackenzie Delta region of Canada (Lantz and Kokelj, 2008), and accelerated thaw slump activity in several areas of western Canada suggest that these features are the dominant drivers of ongoing geomorphic change (Lacelle et al., 2010; Segal et al., 2016). On the Alaskan North Slope, Bowden et al. (2008) identified 34 thermo-erosional features in a 600 km² area near Toolik Lake, two thirds of which have formed in the past 30 years. In northwestern Alaska, the occurrence of thermo-erosional features in the Noatak River Watershed, particularly active layer detachments, increased by more than 200% since the 1980s (Gooseff et al., 2009). The rate of recent permafrost degradation in northern Alaska suggests that 10-30% of Arctic tundra may be affected thermokarst processes, even under modest warming scenarios (Jorgenson et al., 2006).

Retrogressive thaw slumps are a dramatic form of thermo-erosion characterized by a steep headwall and scarp of exposed ground ice, and an adjacent debris flow along a gently sloping slump floor (Lantuit et al., 2012). These features often form following the thaw of relic glacial ice and/or massive ground-ice deposits (Swanson et al., 2014), and are thus commonly found in paraglacial landscapes throughout the Arctic (Segal et al., 2016; Kokelj et al., 2017). Once initiated, the slump headscarp retreats upslope and eventually stops degrading when the exposed ice deposit is covered by sediment (Lantuit et al., 2012). Depending on local soil characteristics, these features can persist for decades to centuries, and are often polycyclic (Lantuit and Pollard, 2008; Kokelj et al., 2009a; Lacelle et al., 2010). The reactivation of pre-existing slump features may be promoted by talik (permafrost-free sediment) enlargement beneath adjacent lakes (Kokelj et al., 2009a). In addition, the insulating effect of snowdrifts created by surface depressions protects the underlying active layer from cold temperatures, which, in conjunction with lower albedo, may act as a positive feedback to further promote polycyclic slump behavior (Osterkamp et al., 2009; Kokelj et al., 2009b). Increased summer
temperatures can alter the thermal regime of soils and initiate thaw slumping, as suggested by evidence for thermo-erosion during the warm Pleistocene-Holocene transition (Mann et al., 2010) and Holocene Thermal Maximum (Burn, 1997; Murton, 2001), as well as recent increases in thaw slump activity in the Arctic coincident with ongoing warming (e.g., Lantz and Kokelj, 2008; Lacelle et al., 2010). Increased moisture and/or timing and intensity of precipitation can also initiate and reactivate retrogressive thaw slumping by increasing slope instability and removing sediments from the slump headwall (Lacelle et al., 2010; Balser et al., 2014, Kokelj et al., 2015; Segal et al., 2016). In addition, wildfires may facilitate thermo-erosion by the rapid removal of insulating organic soils (Liu et al., 2014, Hu et al., 2015, Jones et al., 2015).

Thermo-erosional features such as retrogressive thaw slumps often form along steep slopes surrounding lakes and streams, highlighting an important pathway by which changes to the terrestrial landscape may ultimately impact aquatic systems. Soluble cations sequestered in permafrost and ground ice below the active layer are potential pools of untapped nutrients that may be released during thaw and subsidence. Enhanced mineral sediment mobilization and nutrient loading in nearby lake and stream environments can have dramatic and far-reaching consequences on aquatic productivity and community structure. For example, the release of soluble cations sequestered in permafrost and ground ice can elevate lake water conductivity and alkalinity long after slumps are no longer active (Kojelj and Burn, 2003; Kokelj et al., 2009a; Malone et al., 2013), and even small thermo-erosional features (~2% of the catchment) can have a detectable effect on lake water chemistry (Kokelj et al., 2005). The influx of fine-grained mineral sediments often promotes enhanced water clarity and decreased dissolved organic carbon in the water column of tundra lakes due to adsorption of humics onto fine clay particles (Kokelj et al., 2005; 2009a; Thompson et al., 2008; 2012). In response to increased light conditions and availability of cations such as calcium and magnesium, benthic invertebrates and macrophyte communities flourish on lake margins with shoreline slump features (Mesquita et al., 2010; Moquin et al., 2014). Thaw slumps can also release phosphorus and nitrogen to nearby lakes and streams, which is rapidly taken up by the benthic taxa (Bowden et al., 2008), and slump affected lakes can exhibit greater biological change in species composition through time compared to lakes influenced by regional climate change alone (Thienpoint et al., 2013).

Thermo-erosion may also alter terrestrial ecosystem dynamics and vegetation structure in tundra ecosystems, because thermo-erosional depressions result in changes in topographic relief
that alter near-surface hydrology, snow-pack depth, soil temperature, and nutrient availability. For example, variation in vegetation cover across a chronosequence of Alaskan thermokarst terrains suggests that the abundance of mosses, deciduous shrubs, and evergreen shrubs can increase in response to thermo-erosional disturbance (Schuur et al., 2007). This pattern may be related to enhanced “microtopography” in thermokarst environments, which can result in well-drained microsites preferred by tall shrub species such as willow and birch (Llyod et al., 2003), coupled with the proliferation of mesic bog meadow taxa and Sphagnum spp. in poorly drained depressions (Osterkamp et al., 2009). In addition to hydrological changes, exposures of nonacidic mineral substrates, higher nutrient availability, and warmer soil temperatures in thaw slumps compared to undisturbed terrain may further alter vegetation composition, as suggested by increases in alder cover and productivity in disturbed versus undisturbed sites in northwestern Canada (Lantz et al., 2009). The combined effect of thermokarst microtopography and increased shrub abundance can result in deeper snow drifts that insulate the ground during the winter and further enhance permafrost thaw, suggesting a positive feedback between thermo-erosion and shrub proliferation (Lantz et al., 2009; Sturm et al., 2005). Moreover, thermo-erosional sites colonized by shrubs may act as seed sources within larger areas of undisturbed terrain (Lantz et al., 2009), further promoting the proliferation of deciduous shrub species in the tundra and contributing to the ongoing “greening of the Arctic” (e.g., Epstein, 2012).

Goals of this research

Given the profound environmental changes currently underway in the Arctic, and the potential for novel disturbances in tundra ecoregions in response to ongoing warming, a more thorough understanding the dynamics of tundra disturbance is needed. However, the limited temporal span of observational records in the Arctic, lack of fires in many tundra ecoregions at present, and the spatiotemporal variability of disturbance processes limit the current understanding of tundra disturbance regimes. Paleoecological methods can overcome these limitations by providing an expanded time window to examine changes in disturbance regimes on decadal to millennial timescales, and assess disturbance interactions beyond the scope of observational records. The broad goal of my dissertation research is to examine the patterns, drivers, and interactions of tundra disturbance regimes by using paleoecological methods to reconstruct past fire and thermo-erosional disturbance in the Alaskan tundra.
Although recent large fire events and high-fire years in some Alaskan tundra ecoregions suggest that Arctic fire regimes may be changing in response to ongoing warming (e.g., Hu et al., 2010; 2015), there is limited evidence regarding the long-term patterns of tundra burning by which to assess modern fires. To place modern tundra burning in a broader context of past variability and explore the spatiotemporal patterns of tundra fire regimes (Chapter 2), I used macroscopic charcoal deposited in lake sediments to reconstruct fire history from sites located in three distinct Alaskan ecoregions: the Arctic Foothills, the southern Brooks Range, and the Yukon-Kuskokwim Delta. Sites were chosen from ecoregions characterized by rare fires over the ~60-year observational record and spanning a range of climate conditions and tundra-vegetation types. These lake-sediment archives spanned the past 10,000 to 35,000 years, greatly improving the temporal resolution of tundra fire records in Alaska and providing some of the first long-term perspectives on Alaskan tundra fire history.

Observations over the past ~60 years also suggests that that rate and magnitude of thermo-erosion is increasing in the Arctic (e.g., Jorgenson et al., 2006; Bowden et al., 2008; Lantz et al., 2008; Lacelle et al., 2010; Segal et al., 2016). Despite the increases in both the formation and magnitude of these events in response to anthropogenic climate change, remote sensing of upland areas on the western North Slope revealed a high proportion of degraded thermo-erosional features on the landscape, suggesting this process has occurred frequently in the past in response to natural climate variability (Bowden et al., 2012; Krieger, 2012). Furthermore, retrogressive thaw slumps, which form on the margins of lakes and streams, have been observed not only to reactivate through time, but also to migrate around lake margins (Kokelj et al., 2009b; Lacelle et al., 2010; Lantuit et al., 2012), and it is unclear how long these thermo-erosional processes can remain active. For Chapter 3, I used modern water chemistry and geochemical indicators from lake sediments to examine long-term (i.e., past 6000 years) trends in thermo-erosional activity from Lake NE14 and Perch Lake in the Arctic Foothills ecoregion of the Alaskan North Slope.

If wildfire occurrence and thermo-erosion become more common as warming continues, they may represent a novel and fundamental change in the disturbance regime of tundra ecosystems. Because most tundra regions are characterized by rare fires at present, it is unclear how important increasing fire frequency will be in facilitating thermo-erosion if tundra ecosystems burn more frequently in the future. The Noatak River Watershed is one of the most
flammable tundra ecoregions in Alaska at present (Rocha et al., 2012) and may be an analog for other tundra regions if fires become more common. Previous reconstructions from this region suggest that wildfire activity has been a common occurrence on the landscape for thousands of years, with fire-return intervals ranging from 30-840 years (Higuera et al., 2011b). Moreover, record-high numbers of fires (Higuera et al., 2011a; Rocha et al., 2012) and increasing thermo-erosional activity (Swanson, 2012) has been documented in the Noatak in recent decades. Thus, this region offers a unique opportunity to test the role of fires on facilitating thermo-erosional disturbance. For Chapter 4 of my dissertation, I used a paleoecological approach to compare the timing of past fires events with thermo-erosional episodes at Loon Lake in the Noatak River Watershed of northwestern Alaska. This lake-sediment record spans the past 3000 years, providing a long-term context for testing the role of fire activity in facilitating the initiation of thermo-erosional events and providing compelling insight into the interactions between disturbance regimes in the Arctic.

REFERENCES


Yedoma permafrost amplified by ice wedge thaw: Environmental Research Letters, v. 8, p. 035023.


Zhang, Y., Wolfe, S.A., Morse, P.D., Olthof, I., and Fraser, R.H., 2015, Spatiotemporal impacts of wildfire and climate warming on permafrost across a subarctic region, Canada: Journal of Geophysical Research, v. 120, p. 2338-2356.

CHAPTER 2: SPATIOTEMPORAL PATTERNS OF TUNDRA FIRES: LATE-QUATERNARY CHARCOAL RECORDS FROM ALASKA\textsuperscript{1}

ABSTRACT

Anthropogenic climate change has altered many ecosystem processes in the Arctic tundra and may have resulted in unprecedented fire activity. Evaluating the significance of recent fires requires knowledge from the paleo-fire record because observational data in the Arctic span only several decades, much shorter than the natural fire rotation in Arctic tundra regions. Here we report results of charcoal analysis on lake sediments from four Alaskan lakes to infer the broad spatial and temporal patterns of tundra fire occurrence over the past 35,000 years. Background charcoal accumulation rates are low in all records (range = 0-0.05 pieces cm\textsuperscript{-2} yr\textsuperscript{-1}), suggesting minimal biomass burning across our study areas. Charcoal peak analysis reveals that the mean fire return interval (FRI; years between consecutive fire events) ranged from c. 1650 to 6050 years at our sites, and that the most recent fire events occurred from c. 880 to 7030 years ago, except for the CE 2007 Anaktuvuk River Fire. These mean FRI estimates are longer than the fire rotation periods estimated for the past 63 years in the areas surrounding three of the four study lakes. This result suggests that the frequency of tundra burning was higher over the recent past compared to the late Quaternary in some tundra regions. However, the ranges of FRI estimates from our paleo-fire records overlap with the expected values based on fire-rotation-period estimates from the observational fire data, and the differences are statistically insignificant. Together with previous tundra-fire reconstructions, these data suggest that the rate of tundra burning was spatially variable and that fires were extremely rare in our study areas throughout the late Quaternary. Given the rarity of tundra burning over multiple millennia in our study areas and the pronounced effects of fire on tundra ecosystem processes such as carbon cycling, dramatic tundra ecosystem changes are expected if anthropogenic climate change leads to more frequent tundra fires.

INTRODUCTION

The tundra biome occupies some of the coldest regions on Earth and is thus characterized by low biomass compared to other ecosystems. Despite low productivity in tundra ecosystems, circumpolar Arctic regions account for approximately 50% of all belowground soil organic carbon (Schuur et al., 2008; Grosse et al., 2011), in part because low decomposition rates and infrequent burning allow for carbon accumulation over millennia. In Alaska, observational records show that fire has been rare in the majority of tundra ecoregions during the past 60 years (Rocha et al., 2012). However, anthropogenic climate change may have increased the rate of tundra burning. For example, in Common Era (CE) 2007, the Anaktuvuk River Fire (ARF) burned approximately 1000 km$^2$, doubling the total area burned on the Alaskan North Slope since CE 1950 (Jones et al., 2009; Mack et al., 2011). The Noatak River Watershed, a tundra region in northwestern Alaska that has historically burned more frequently than the North Slope, also experienced an increase in area burned over the past several decades (Rocha et al., 2012) and a record high number of fires in CE 2010 (AICC, 1943–2013). With anticipated acceleration of anthropogenic climate change in the Arctic, fires may become increasingly important in tundra regions that rarely burn at present.

Tundra fires can dramatically impact a variety of ecosystem processes. For example, the ARF released an amount of carbon comparable to the net carbon sink of the entire Arctic tundra biome in a typical year in the latter part of the 20th century (Mack et al., 2011). Decreased organic soil thickness and moss cover following the fire resulted in changes to the ground thermal regime, including increased permafrost thaw depth and higher soil temperatures (Rocha and Shaver, 2011). Enhanced microbial activity and access to deeper soil layers associated with permafrost thaw can further increase the release of tundra-soil carbon to the atmosphere over decadal timescales. Thus increased tundra burning in response to anthropogenic climate change may lead to pronounced ecosystem changes.

The brevity of the observational fire record makes it difficult to characterize the variability and drivers of tundra fire regimes. Therefore, fire-history reconstructions from lake-sediment charcoal analysis provide key information on the long-term dynamics of tundra burning and a necessary context to assess recent changes. For example, paleofire data reveal that the area within the ARF had not experienced fire in at least 5000 years (Hu et al., 2010). In contrast, paleorecords from the Noatak River Watershed suggest frequent tundra burning over the past
2000 years, with mean fire return intervals (FRI, the time interval between consecutive fires) comparable to those in modern-day boreal forests (c. 100-300 years; Higuera et al., 2011). Paleorecords also reveal vegetation-mediated responses of tundra fire regimes to climate change, such as an increase in fire frequency in north-central Alaska in association with the expansion of shrubs in the tundra vegetation of the last glacial/interglacial transition (Higuera et al., 2009). However, existing paleorecords of tundra burning are restricted to a few sites (Fig. 2.1), and we know little about the patterns and drivers of tundra burning elsewhere. To address this limitation, and to place modern fire regimes in a broader context of past variability, we conducted charcoal analysis of sediment cores from four lakes in Alaska. The results allow us to examine the spatiotemporal patterns of fire regimes over the late Quaternary and provide a context of natural fire-regime variability for assessing recent tundra burning.

STUDY SITES

Our four study sites are located in three tundra ecoregions of Alaska that are characterized by a paucity of fires in the observational record and that span a range of climate conditions and tundra-vegetation types. Ecoregion classification and descriptions follow Nowacki et al. (2001), modified to delineate the Brooks Range Transition zone between boreal and tundra vegetation as a distinct ecoregion (Fig. 2.1). For modern climate near each site, June-August (JJA) average temperature and total precipitation were estimated within a 5-km radius around each lake (Table 2.1), using data from the Parameter-elevation Regression on Independent Slopes Model (PRISM Climate Group, 2012) spanning 1971–2000 (data downloaded from SNAP, 2014).

Perch Lake (68.94° N, 150.50° W) and Upper Capsul Lake (68.63° N, 149.41° W) are small kettle basins located in the Brooks Range Foothills ecoregion (hereafter referred to as the North Slope; Fig. 2.1), which is characterized by gently rolling hills, narrow alluvial valleys, and surficial deposits comprised primarily of glacial moraines, outwash, and alluvial materials. Mean JJA temperature in this area is 10.5 ± 0.4 °C, and total JJA precipitation is 144 ± 44 mm (1971-2000 mean and standard deviation; Table 2.1). Soils in the region feature continuous permafrost overlain by organic-rich horizons. Vegetation is dominated by mixed shrub-sedge tussock tundra, interspersed with willow thickets along rivers and small drainages. Perch Lake lies within
the Anaktuvuk River Fire (ARF), and Upper Capsule Lake is approximately 50 km to the southeast of Perch Lake and 40 km from the southernmost portion of the ARF (Fig. 2.1).

Keche Lake (68.02° N, 146.92° W) lies in the southeastern portion of the Brooks Range ecoregion. Sedimentary and metamorphic deposits dominate this steep mountainous terrain. Mean JJA temperature and total JJA precipitation in the area are 10.8 ± 0.5 °C and 142 ± 14 mm, respectively. The modern vegetation around Keche Lake is forest tundra, in the transition zone between tundra and boreal forest, as defined by the Circumpolar Arctic Vegetation Map (CAVM Team, 2003; Walker et al., 2005). The area is designated as the Brooks Range Transition (Fig. 2.1), and the lake is approximately 200 m below treeline with stands of *Picea glauca* (white spruce) in the watershed. The early-Holocene vegetation in this area was shrub tundra based on the regional pollen dataset (Anderson and Brubaker, 1994).

Tungak Lake (61.43° N, 164.20° W) is located in the broad Yukon-Kuskokwim Delta ecoregion of southwestern Alaska. Mean JJA temperature and total JJA precipitation in the Tungak Lake area are 12.3 ± 0.1 °C and 169 ± 2 mm, respectively. The regional landscape is characterized by shallow organic soils, discontinuous permafrost, and abundant thermokarst lakes. Tungak Lake is located in an isolated area of low-shrub tundra surrounded by low-shrub wetlands.

**MATERIALS AND METHODS**

Two overlapping sediment cores were obtained from the deepest portion of Keche and Tungak lakes in the summers of 2007 and 2012, respectively. Perch Lake was first cored in 2008, and charcoal analysis on the core was conducted to infer fire history of the past 5000 years (Hu et al., 2010). For this study, we include additional data from deeper sediments obtained in 2011, extending the fire record to the past c. 9500 years. The sediment cores from Upper Capsule Lake were obtained in 1997 for pollen analysis (Oswald et al., 2003). At each lake, a polycarbonate tube fitted with a piston was used to retrieve an intact sediment-water interface and the uppermost sediments, and a modified Livingstone piston corer (Wright et al., 1984) was used to obtain deeper sediments. The top 5-20 cm of unconsolidated surface sediments were extruded at 0.5-cm resolution in the field, and the remaining sections were split lengthwise in the laboratory. Overlapping cores were correlated based on visible stratigraphic transitions and magnetic susceptibility.
Chronologies are based on $^{210}\text{Pb}$ analysis on bulk sediments (except at Upper Capsule Lake, where no $^{210}\text{Pb}$ analysis was performed) and AMS $^{14}\text{C}$ analysis on terrestrial macrofossils (Fig. 2.2; Table A.1). Preparation of $^{210}\text{Pb}$ samples followed Eakins and Morrison (1978), and activity was measured with an Ortec Octête Plus alpha spectrometer at the University of Illinois. We used a constant-rate-of-supply (CRS) model adapted from Binford (1990) to estimate $^{210}\text{Pb}$-based sample ages. For $^{14}\text{C}$ measurements, terrestrial macrofossils were treated with an acid-base-acid procedure (Oswald et al., 2005) and submitted to Lawrence Livermore National Laboratory (Livermore, CA) or INSTAAR Radiocarbon Laboratory (Boulder, CO). All $^{14}\text{C}$ ages were calibrated to years before CE 1950 (cal BP) using the IntCal 09 dataset in CALIB v6.1.0 (Stuiver and Reimer, 1993; Reimer et al., 2009). A thick tephra was visible in the sediments of Tungak Lake spanning 5-38 cm. We assume that this tephra was deposited during the Aniakchak eruption, which was widespread in the region with a well-constrained age of $3.7 \pm 0.2$ kcal BP (Begét et al., 1992; Kaufman et al., 2012). We adjusted the depth of the sediment core by assuming that the tephra deposited instantaneously. Age models were developed by fitting a weighted cubic smoothing spline through all ages, and confidence intervals were estimated with bootstrap resampling using the MCAgeDepth program (2009) (Higuera et al., 2009).

For charcoal analysis, 0.5-2.0 cm$^3$ subsamples were taken from continuous 0.25-1.0 cm core slices. Sediments were freeze-dried overnight, immersed in 5 ml of bleach and 5 ml of 10% sodium metaphosphate for approximately 20 h, and then washed through a 125 μm sieve. Charcoal particles >125 μm were enumerated under a dissecting microscope (10-40x magnification). Because charcoal counts are low at all of our sites, count data from adjacent samples were aggregated to obtain a final sampling resolution of 0.5-1.0 cm and volume of 2-4 cm$^3$. Charcoal concentrations (pieces cm$^3$) were multiplied by the sediment accumulation rate (cm year$^{-1}$) to calculate charcoal accumulation rates (CHAR, pieces cm$^{-2}$ year$^{-1}$).

Although tundra fires consume lower biomass than fires in forest ecosystems, previous studies have shown that charcoal production is sufficient for reliable detection of local fires in lake-sediment records (Hu et al., 2010; Higuera et al., 2011). We infer local fires (within 500-1000 m of each lake; Higuera et al., 2007; 2011) from our CHAR records using CharAnalysis v1.1 program (2013), modified as described below. Prior to statistical analyses, charcoal samples were interpolated to the median sample resolution of each record (Table 2.2) to account for unequal sampling from variable sediment accumulation rates. Low and zero charcoal counts
were prevalent in all records. To guard against interpretation of fluctuations based on small differences between samples, we used a wide time-window to estimate the low frequency component and limit our interpretation of “background” CHAR to broad trends in the data. Background trends in each interpolated CHAR record were estimated using a Lowess smoother (Cleveland, 1979) with a 3000-year time window. A detrended series was created by subtracting this low frequency trend from the interpolated CHAR series. The method commonly used for establishing the threshold for charcoal peak detection is based on the assumption that the noise component is normally distributed around the background trend (Higuera et al., 2010). However, this assumption is poorly met in our records because of the prevalence of CHAR values of exactly zero. Thus, we used a zero-inflated gamma (ZIG) distribution to separate the detrended series into “noise” and “peak” components, specifying the 99th percentile of the distribution as the global threshold for each record. The noise component is assumed to reflect random variability, such as charcoal deposition from distant fires and/or local depositional processes, and the peak component is used to identify fire events within the interpolated sample (e.g. Gavin et al., 2003; Lynch et al., 2004; Higuera et al., 2007).

We used a minimum count screening to remove peak identification that could arise from statistical noise associated with low charcoal particle counts (Gavin et al., 2006). If the charcoal count for a peak sample had a >15% probability of being drawn from the same Poisson distribution as the lowest non-peak count within the previous 1500 years, the peak was rejected. After thresholds were determined, we calculated a signal-to-noise index (SNI) to evaluate the suitability of our records for peak detection (Kelly et al., 2011). The identified charcoal peaks were interpreted as fire events, and fire-return intervals (FRIs) were calculated as years between individual fire events.

To place our fire history reconstructions in the context of fires on the modern landscape, we calculated the fire-rotation period (FRP, also termed fire cycle; Johnson and Gutsell, 1994) for each site. The FRP value calculated from spatially explicit data of fire observations is equivalent to the mean FRI calculated from temporal variations in fire occurrence for any point on the landscape (Johnson and Gutsell, 1994), and thus modern FRP can be compared to paleo-inferred mean FRI (Kelly et al., 2013). We defined the FRP at each lake as \( FRP = \frac{t}{\sum_{i=1}^{n} a_i/A} \), where \( t \) is the temporal span of the historical fire record (CE 1950-2013, 63 years), \( a_i \) is the area (km\(^2\)) burned by fire \( i \), \( n \) is the total number of fires (range = 2 to 19), and \( A \) is the vegetated area.
(km²) within the 100 km buffer (Baker, 2009). We found that site-specific FRP calculations were generally stable for radii between 60 and 140 km, suggesting the 100 km radius is an appropriate area to characterize the modern fire regime. To obtain the vegetated area within each buffer, we subtracted barren and open water land cover classes (defined by the National Landcover Database vegetation survey from 2006 (NLDC, 2006) and the North American Land Change Monitoring System (NALCMS, 2005)) from the total area, based on the rationale that these cover types do not have burnable fuels. We also removed any fire perimeters that were described as human-caused. We present each FRP with the 95% quantile range from an exponential distribution with mean equal to the FRP. These bounds represent the likely range of individual FRIs expected for a fire regime defined by the estimated FRP.

RESULTS AND DISCUSSION

Chronologies

The age-depth models of our four sediment records were based on a total of 56²¹⁰Pb-estimated ages, 41 calibrated ¹⁴C ages, and one tephra-based age (Fig. 2.2; Table A.1). The chronology for Upper Capsule Lake follows Oswald et al. (2003). The ¹⁴C ages for the other three lakes are all in chronological order with the exception of the age at 107 cm in the Keche Lake core. This age was excluded from chronological modeling as it was considered too old based on surrounding dates; the dated material likely had resided in watershed soils before deposition in the lake, a common ¹⁴C-dating problem for Arcto-boreal sediments (Oswald et al. 2005). The density of ¹⁴C dates varies across the four sites; for example, the Tungak Lake chronology is constrained by only five ¹⁴C ages for the past 35,500 years, whereas the Perch Lake chronology is constrained by 10 ¹⁴C ages for the past 9500 years.

Sedimentation rates are relatively low across all four sites. Based on the age-depth model, the Perch Lake record spans the past c. 9500 years, with an average sedimentation rate (± standard deviation) of 0.04 ± 0.08 cm year⁻¹ (Table 2.1). The Upper Capsule record spans the past c. 12,100 years, with an average sedimentation rate of 0.03 ± 0.01 cm year⁻¹. The Keche Lake sediment core has a modeled basal age of 11.5 kcal BP and an average sedimentation rate of 0.03 ± 0.03 cm year⁻¹. Tungak Lake has the oldest sediment sequence in this study, spanning the past c. 35,500 years. The sedimentation rate changes at 38 cm from 0.03 ± 0.04 cm year⁻¹ between 12.0 and 35.5 kcal BP to 0.01 ± 0.03 cm year⁻¹ after 12.0 kcal BP. These relatively low
sedimentation rates did not present a problem for the identification of local fires because of the rarity of fires across all four sites (see below).

**Spatial and temporal patterns of fire occurrence**

The two charcoal records from the North Slope both exhibit low background CHAR (mean = 0.008 pieces cm\(^{-2}\) yr\(^{-1}\) for Perch and Upper Capsule) (Fig. 2.3B), suggesting minimal biomass burned in the region over the past c. 12,000 years. For comparison, background CHAR values have means of 0.340 pieces cm\(^{-2}\) yr\(^{-1}\) in boreal fire records from interior Alaska (Kelly et al., 2013) and 0.012 pieces cm\(^{-2}\) yr\(^{-1}\) in the tundra fire records of the Noatak River Watershed (Higuera et al., 2011). Charcoal peak analysis identified only three local fires in the Perch Lake record, at 9.4 kcal BP, 6.5 kcal BP, and CE 2007 (Fig. 2.3C). This result confirms the published finding from this site that the ARF in CE 2007 was unprecedented in the past 5000 years (Hu et al., 2010), and extends the uniqueness of this fire event to the past 6500 years. Around Upper Capsule Lake, only one fire occurred during the past c. 12,000 years. This fire is dated at c. 6.45 kcal BP (Fig. 2.3C), coincident with the Perch Lake fire at c. 6.48 kcal BP. Given the local origin of macroscopic charcoal peaks (0.5–1.0 km; Higuera et al., 2007) and the distance between the two lakes (~50 km), it is unlikely that a fire event at one site resulted in a charcoal peak at the other site. Instead, the presence of a prominent charcoal peak at 6.5 kcal BP in both records likely represents a single large fire that burned across both sites. Because the magnitude of charcoal peaks reflects, in part, the amount of burned biomass (e.g. Whitlock et al., 2006; Higuera et al., 2009), the higher CHAR peak at 6.5 kcal BP at Perch Lake suggests that the amount of biomass consumed in the fire at 6.5 kcal BP was greater than that of the ARF. Alternatively, two separate but similarly-timed events may have occurred in the watersheds of these lakes. Because vegetation has changed little over the past c. 7000 years (Oswald et al., 2003), this fire event suggests that climate and/or ignition constraints on burning were relaxed during this time, perhaps as least as warm and dry as the anomalous climatic conditions that facilitated the ARF (Hu et al., 2010). However, we cannot verify this interpretation because no suitable paleoclimate record with seasonal resolution is available from the region (Oswald et al., 2014).

The Keche Lake record spans several millennia during the early Holocene when tundra vegetation dominated the regional landscape and climate was generally cooler and drier than
modern (Anderson and Brubaker, 1994; Kaufman et al., in review, 2015; Clegg et al., 2011). Background CHAR is exceptionally low, with a mean of 0.0007 pieces cm\(^{-2}\) yr\(^{-1}\) from 11.4 to 8.8 kcal BP (Fig. 2.3B), suggesting little burning on the early-Holocene landscape of the region. No local fires occurred around Keche Lake during this period. Background CHAR increases to a mean of 0.02 pieces cm\(^{-2}\) yr\(^{-1}\) between 8.8 and 4.5 kcal BP (Fig. 2.3B), suggesting an increase in regional burning coincident with the development of a forest-tundra ecotone in the Alaskan interior with sparse stands of *Picea glauca* (white spruce) near Keche Lake by c. 9.0 kcal BP (Anderson and Brubaker, 1994). Local fires were more frequent at Keche Lake between 8.8 and 4.5 kcal BP than during the early and late Holocene, with five events at 7.6, 7.4, 6.5, 5.1, and 4.6 kcal BP (Fig. 2.3C). This change implies that tundra burning was limited either by cooler summer temperatures in the early Holocene or by a lack of biomass, given that the early Holocene was drier than the middle Holocene in the Alaskan interior (Kaufman et al., in review, 2015), which should have favored burning. It is possible that the lack of fire in the early Holocene resulted from locally moist conditions in summer, as suggested by peatland expansion that began c. 11.2-10.7 kcal BP at a site ~20 km to the northwest of Keche Lake (Jones and Yu, 2010). Background CHAR decreases to 0.004 pieces cm\(^{-2}\) yr\(^{-1}\) from 4.5 kcal BP to present, and only one fire event occurred during the past c. 4500 years near Keche Lake (Fig. 2.3). In contrast, area burned and fire frequency increased after 4.0 kcal BP in the boreal forests of interior Alaska (Higuera et al., 2009; Hu et al., 2006; Kelly et al., 2013). This contrast can be attributed to the development of flammable forests dominated by *P. mariana* (black spruce) in the lowlands of interior Alaska (Higuera et al., 2009; Kelly et al., 2013), and the absence of this species in upland treeline areas around Keche Lake. The low fire frequency at Keche Lake after 4.5 kcal BP may have resulted from decreasing summer temperatures associated with late-Holocene neoglacialiation (e.g. Barclay et al., 2009; Clegg et al., 2011; Badding et al., 2013).

The Tungak Lake record is the longest fire-history reconstruction from Alaska. Throughout the past c. 35,000 years, low background CHAR in this record (mean = 0.002 pieces cm\(^{-2}\) yr\(^{-1}\); Fig. 2.3B) suggests little burning, likely resulting from a combination of cold and sometimes arid conditions that limited biomass in the widespread graminoid-herb tundra of the region (Ager et al., 2003; Kurek et al., 2009). Only five local fires are identified over the past 35,000 years (Fig. 2.3C). Between 25.5 and 14.0 kcal BP, background CHAR is generally higher than the remainder of the record, and peak analysis shows four local fire events at 25.4, 17.8,
16.7, and 14.4 kcal BP. During this period, the modern-day coastal regions of Alaska experienced greater continentality because of lower sea levels, resulting in more arid conditions than today (e.g. Alfimov and Berman, 2001; Kurek et al., 2009). Such conditions may have relaxed the climatic constraints on tundra burning, leading to more frequent fire events between 25.5 and 14.0 kcal BP compared to the Holocene (i.e., after 11.7 kcal BP). During the Holocene, background CHAR is lower than during the glacial period, and peak analysis suggests only one fire event, at 7.0 kcal BP. Thus fire activity decreased during the Holocene, probably as a result of increased effective moisture in the region despite increased tundra biomass compared to the glacial period (Ager, 2003; Hu et al., 1995).

Overall, the most striking feature of our records is that these tundra regions have generally persisted as rare-fire systems for many millennia. With the exception of the ARF on the North Slope, the most recent fire events at these sites occurred from 882 to 7031 years ago (Table 2.2; Fig. 2.3C). These results stand in stark contrast with the paleo-fire data from the tundra ecosystems of the Noatak River Watershed. In that area, mean FRIs ranged from 135 to 309 years based on charcoal records of the past 2000 years from several lakes, comparable to mean FRIs from modern boreal forests in Alaska (Higuera et al., 2011). In north-central Alaska, Higuera et al. (2008) documented frequent tundra fires during the late-glacial and early-Holocene period between 14 and 10 kcal BP, a finding supported by a microscopic charcoal record from central Alaska (Tinner et al., 2006). However, evaluating the spatial extent of this finding has not been possible because of the general lack of paleofire data from this period. Three of our four charcoal records span the entirety (Tungak Lake) or a portion of this period (Upper Capsule and Keche lakes). These new records show no enhanced fire activity during this period; only one local fire occurred near this time period (at 14.4 kcal BP at Tungak Lake), and regional biomass burning was consistently low across all three sites between 14 and 10 kcal BP (Fig. 2.3). Together with the previous fire-history reconstructions, our data suggest that the rate of tundra burning was spatially variable throughout the late Quaternary.

**Recent tundra burning in the context of paleo-fire records**

Our paleo-fire records can provide a context for comparison with modern tundra fire regimes by invoking the statistical equivalency of the mean FRI and the modern fire rotation period (FRP; Johnson and Gutsell, 1994). We compared our paleo-based mean FRI estimates
with the FRP values calculated from spatially explicit data of modern fire observations spanning the past 63 years. Across our four sites, modern FRPs within 100 km of each lake ranged from 771 years (Keche Lake) to 7560 years (Tungak Lake), with intermediate values at Perch (1909 years) and Upper Capsule (2070 years) lakes (Table 2). Although the rarity of tundra burning makes these estimates highly uncertain, the spatial patterns are similar to those in our paleo-fire records, suggesting that the differences in tundra burning across our study sites have been present over long timescales (Fig. 2.4). These variations are largely related to spatial heterogeneity in climate (Young et al., 2013), consistent with the finding that summer temperature and precipitation explained 95% of the interannual variability in area burned in Alaskan tundra (Hu et al., 2010).

Our paleo-fire analyses underestimate the true mean FRI because the oldest and most recent fire events in each record only provide minimum FRI estimates (i.e., censored intervals; Fig. 2.4). Despite this underestimation, at three of our four study sites, the mean FRI estimates from the paleo-fire records are longer than the FRP estimates based on recent fires. Specifically, the mean FRI (range) estimates of 4730 (2924-6536) years at Perch Lake and 6045 (>5590->6500) years at Upper Capsule Lake are much longer than the FRP (95% quantile range) estimates of 1909 (48-7042) and 2070 (52-7636) years for these lakes, respectively (Fig. 2.4; Table 2.2). Likewise, the mean FRI estimate of 1648 (144-3906) years at Keche Lake is longer than the modern FRP estimate of 771 (19-2844) years. The exception is Tungak Lake where the mean FRI of 5904 (1157->9968) is shorter than the FRP of 7560 (191-27,888) years. This shorter mean FRI reflects more frequent burning between 25.5 and 14.0 kcal BP, when the region was probably more arid than during the Holocene. The FRP of 7560 years is similar to the most recent individual FRI of >7031 years at that site (Table 2.2). Thus our analysis suggests that the frequency of tundra burning was higher over the past 63 years at three of our four sites compared to the late Quaternary. This inference suggests elevated fire activity in some tundra regions at present, possibly as a result of anthropogenic climate change.

However, the above comparison is inconclusive, because the range of individual FRIs at each site generally falls within the broad range of FRIs that can be expected to arise by chance, as defined by the 95% quantile range around the FRP estimates (Fig. 2.4). Furthermore, the ranges of individual FRIs from the paleorecords are likely influenced by past climate and vegetation conditions that differed from modern conditions, and the estimates of both mean FRI
and FRP are uncertain due to the rarity of tundra fires and thus high sensitivity to individual fire events. For example, without the ARF, the FRP in the tundra region of Perch and Upper Capsule lakes would be >100,000 years, which is much longer than the FRP estimates of Perch (1909 years) and Upper Capsule (2070 years) lakes with the inclusion of the ARF. Thus quantitative comparisons between the mean FRI estimated from our paleorecords and the FRP estimates from historic observational records inherently contain a large amount of uncertainty.

Paleorecords provide critical information regarding natural variability and thus play an important role in assessing potential anthropogenic changes in climate and ecosystems. Our results comparing paleo-inferred mean FRI and observed modern FRP illustrate an important limitation of using paleo-fire records in systems that rarely burn to quantitatively assess whether recent burning is beyond the range of natural variability. This limitation is applicable to other situations where the events of interest have rarely occurred in the past. One way to circumvent this limitation is to increase the spatial density of paleorecords and pool the data of detected past events to improve statistical power (Whitlock et al., 2010). Several paleofire studies have demonstrated the value of increasing the spatial density of sampling for bolstering the confidence in inferences about recent changes (Marlon et al., 2012; Kelly et al., 2013).

The utility of paleorecords with rare events may improve if the frequency of such events increases markedly in the future. To examine how much burning would be required to shift the fire regime unequivocally beyond the range of past FRIs, we considered hypothetical scenarios of future burning around Perch Lake, which has two well-constrained (i.e., uncensored) FRIs in the Holocene paleo-fire record. We calculated the FRP for CE 2050 assuming the addition of one to four fires of the ARF size (~1000 km²) within the 100-km radius around Perch Lake (Fig. 2.5). One additional ARF-sized fire would shift the Perch Lake FRP estimate to 1502 years. This FRP is much shorter than the mean FRI of 4730 years based on the paleo-fire record from Perch Lake, and the most recent FRI at the lake (6536 years) is well outside the 95% quantile range indicated by this updated FRP estimate. Thus, the addition of a single ARF-sized fire within the 100-km region surrounding Perch Lake would offer compelling evidence that the modern fire regime represents a significant increase in fire activity. The occurrence of additional large fire events would further decrease the FRP and strengthen confidence in that estimate, making it increasingly difficult to accept the null hypothesis that the modern fire regime is consistent with past variability. Thus, although the rarity of fire events makes it uncertain as to whether modern
fire regimes differ from the past, the uncertainty will be reduced as the observational record
grows, especially if a dramatic fire-regime shift is indeed underway. Increasing the spatial
density of paleorecords to refine our understanding of past variability would enhance the rigor of
testing whether or not recent and future changes in fire regimes are truly unprecedented.

TABLES

Table 2.1: Lake characteristics for the four study sites. June–August (JJA) climatology is from
PRISM-derived data spanning 1971-2000, summarized over an approximate radius of 5-km
around each lake (representing 20-21 PRISM pixels). Circumpolar Arctic Vegetation Map
(CAVM) landcover classification is based on Walker et al. (2005). ‘Boreal transition’ indicates
that the site is outside of the CAVM classification.

<table>
<thead>
<tr>
<th>Characteristic</th>
<th>Site</th>
<th>Perch</th>
<th>Upper Capsule</th>
<th>Keche</th>
<th>Tungak</th>
</tr>
</thead>
<tbody>
<tr>
<td>Site</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Latitude</td>
<td>68° 56’ 29.4” N</td>
<td>68° 37’ 43.0” N</td>
<td>68° 1’ 2.8” N</td>
<td>61° 25’ 37.9” N</td>
<td></td>
</tr>
<tr>
<td>Longitude</td>
<td>150° 29’ 57.7” W</td>
<td>149° 24’ 48.7” W</td>
<td>146° 55’ 25.7” W</td>
<td>164° 12’ 2.2” W</td>
<td></td>
</tr>
<tr>
<td>Elevation (m a.s.l.)</td>
<td>400</td>
<td>800</td>
<td>740</td>
<td>25</td>
<td></td>
</tr>
<tr>
<td>Surface area (ha)</td>
<td>14.0</td>
<td>1.1</td>
<td>80.2</td>
<td>117.0</td>
<td></td>
</tr>
<tr>
<td>Max. water depth (m)</td>
<td>12.6</td>
<td>5.7</td>
<td>15</td>
<td>15.4</td>
<td></td>
</tr>
<tr>
<td>Coring water depth (m)</td>
<td>12.6</td>
<td>5.7</td>
<td>14.5</td>
<td>14.8</td>
<td></td>
</tr>
<tr>
<td>CAVM landcover class</td>
<td>Tussock-sedge, dwarf-shrub tundra</td>
<td>Tussock-sedge, dwarf-shrub tundra</td>
<td>Boreal transition (near treeline)</td>
<td>Low-shrub tundra</td>
<td></td>
</tr>
<tr>
<td>JJA temperature (°C)</td>
<td>10.9 ± 0.04</td>
<td>10.0 ± 0.1</td>
<td>10.8 ± 0.5</td>
<td>12.3 ± 0.1</td>
<td></td>
</tr>
<tr>
<td>JJA total precip. (mm)</td>
<td>101 ± 5</td>
<td>187 ± 7</td>
<td>142 ± 14</td>
<td>169 ± 2</td>
<td></td>
</tr>
<tr>
<td>Record</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Core length (cm)</td>
<td>209.5</td>
<td>329</td>
<td>321</td>
<td>353.5</td>
<td></td>
</tr>
<tr>
<td>Coring year (cal yr BP)</td>
<td>-58</td>
<td>-47</td>
<td>-57</td>
<td>-62</td>
<td></td>
</tr>
<tr>
<td>Basal age (cal yr BP)</td>
<td>9460</td>
<td>12100</td>
<td>11480</td>
<td>35430</td>
<td></td>
</tr>
<tr>
<td>Sed. rate (cm yr⁻¹)</td>
<td>0.045 ± 0.080</td>
<td>0.031 ± 0.011</td>
<td>0.030 ± 0.027</td>
<td>0.030 ± 0.039</td>
<td></td>
</tr>
</tbody>
</table>
Table 2.2: Results of charcoal analysis and modern fire rotation period (FRP) from all study sites. Fire-return intervals (FRIs) are given as the range, mean, and most recent FRI (‘>’ indicates that there is no modern or previous fire in the record to constrain the interval). Modern FRPs are calculated for the vegetated areas within a 100-km radius around each lake, with human-caused fires excluded from analysis. For each FRP, a 95% quantile range of expected FRIs is calculated assuming an exponential distribution with the mean equal to the FRP.

<table>
<thead>
<tr>
<th>Charcoal Analysis</th>
<th>Site</th>
<th>Perch</th>
<th>Upper Capsule</th>
<th>Keche</th>
<th>Tungak</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean sample resolution (yr sample⁻¹)</td>
<td></td>
<td>22.7 ± 13.1</td>
<td>37.1 ± 14.1</td>
<td>9.3 ± 1.4</td>
<td>50.3 ± 49.4</td>
</tr>
<tr>
<td>Average (range) charcoal count (pieces cm⁻³)</td>
<td></td>
<td>2.1 (0-114)</td>
<td>0.5 (0-11)</td>
<td>0.7 (0-85)</td>
<td>0.7 (0-25)</td>
</tr>
<tr>
<td>Average (range) charcoal conc. (pieces cm⁻³)</td>
<td></td>
<td>0.56 (0-28.5)</td>
<td>0.14 (0-2.75)</td>
<td>0.29 (0-34.0)</td>
<td>0.19 (0-6.25)</td>
</tr>
<tr>
<td>Interpolation interval (yr)</td>
<td></td>
<td>43</td>
<td>65</td>
<td>18</td>
<td>89</td>
</tr>
<tr>
<td>#of peaks identified as fires</td>
<td></td>
<td>3</td>
<td>1</td>
<td>6</td>
<td>5</td>
</tr>
<tr>
<td>Range of FRIs (yr)</td>
<td></td>
<td>2924 - 6536</td>
<td>&gt;5590 - 6500</td>
<td>144 - 3906</td>
<td>1157 - &gt;9968</td>
</tr>
<tr>
<td>Mean FRI (yr)</td>
<td></td>
<td>4730</td>
<td>6045</td>
<td>1648</td>
<td>5904</td>
</tr>
<tr>
<td>Most recent FRI (yr)</td>
<td></td>
<td>6536</td>
<td>&gt;6500</td>
<td>&gt;882</td>
<td>&gt;7031</td>
</tr>
</tbody>
</table>

Modern Fire Rotation Period (FRP)

| Burnable area within 100-km radius (km²)                | 27809         | 21976 | 21666         | 13814           |
| #of observed fires 1950-2013                            | 3             | 2     | 19            | 12              |
| Area burned 1950-2013 (km²)                            | 932.1         | 679.4 | 1798.5        | 116.9           |
| FRP (yr)                                               | 1909          | 2070  | 771           | 7560            |
| 95% quantile range, expected FRIs (yr)                 | 48-7042       | 52-7636 | 19-2844       | 191-27888       |

30
Figure 2.1: Map of tundra ecoregions (modified after Nowacki et al., 2001) with study sites (shown with 100-km buffers) and locations of previous tundra fire-history reconstructions. CE 1950-2013 fire perimeters from http://fire.ak.blm.gov.
Figure 2.2: Age-depth relationships for all sites modeled with a cubic spline and presented with 95% confidence intervals.
Figure 2.3: Charcoal records and peak analysis. A) Raw and interpolated charcoal accumulation rates. B) Background (i.e., low-frequency variability) charcoal accumulation rates. C) Charcoal peak identification.
Figure 2.4: Fire rotation period (FRP) estimated for the 100-km buffer around each site, and individual and mean paleo-based fire-return intervals (FRIs) from the sediment charcoal records. 95% quantile ranges represent expected values for individual FRIs, based on the estimated FRP. Censored FRIs are individual FRIs with no modern or previous fire to constrain the interval.
Figure 2.5: Perch Lake paleo fire-return intervals (FRIs), modern fire rotation period (FRP), and FRPs for CE 2050 assuming one to four additional large fires within the 100-km radius around the lake. ARF-sized fires = 919.7 km$^2$, which is the total vegetated area burned during the Anaktuvuk River Fire in a 100-km radius around Perch Lake. Mean and individual FRIs from Perch Lake on the far left (symbols same as Figure 2.4). FRP estimates shown with 95% quantile range of expected FRIs assuming an exponential distribution with the mean equal to the FRP.
REFERENCES


CHAPTER 3: MULTIPLE THERMO-EROSIONAL EPISODES DURING THE PAST SIX MILLENNIA: IMPLICATIONS FOR THE RESPONSE OF ARCTIC PERMAFROST TO CLIMATE CHANGE ²

ABSTRACT

Anthropogenic warming may promote rapid permafrost thaw in the Arctic and alter the global carbon cycle. Although several studies suggest increased thermo-erosion as a result of recent warming, a long-term context is necessary to assess the linkages of thermokarst processes with climate variability. We analyzed sediment cores from two lakes on the Alaskan North Slope, one with (Lake NE14) and one without (Perch Lake) watershed thermo-erosion. Distinct geochemical and lithological characteristics provide evidence for sedimentary input from carbonate-rich permafrost soils associated with past retrogressive thaw slumping at NE14 but not at Perch. These characteristics include increases in Ca:Sr, Ca:K, carbonate:[feldspar + clay minerals], %CaCO₃, and δ¹³C, and decreases in ^⁸⁷Sr:^⁸⁶Sr. At least ten episodes of thermo-erosion occurred over the past 6000 years at NE14. Most of these episodes coincided with periods of elevated summer temperatures, but moisture variation and geomorphic factors likely played a role in driving their occurrence. Our results suggest that positive feedbacks facilitate reactivation of thermo-erosion in ice-rich terrain, adding to the growing body of evidence that these Arctic landscapes are unstable in a changing climate.

INTRODUCTION

Anthropogenic climate warming in the Arctic (Kaufman et al., 2009) is expected to accelerate because of complex feedbacks (Serreze and Barry, 2011). In particular, permafrost soils contain twice as much carbon as the modern atmosphere (Tarnocai et al., 2009), and thus climate-driven permafrost thaw may elevate atmospheric CO₂ concentration and amplify global warming (Schuur et al., 2015). Thermo-erosion is the thaw-induced subsidence and mass movement of ice-rich soils that facilitate rapid carbon release by exposing ancient carbon to decomposition (Cory et al., 2013) and losses via riverine transport (Schreiner et al., 2014). Thermo-erosion can also stimulate carbon sequestration by enhancing nutrient availability and

² This chapter is published as: Chipman, M.L., Kling, G.W., Lundstrom C.C., and F.S. Hu (2016) Multiple thermo-erosional episodes during the past six millennia: Implications for the response of Arctic permafrost to climate change. Geology, 44(6), 439-442.
plant productivity (Schuur et al., 2007). Current models of carbon emissions from permafrost thaw do not account for the complexity of thermo-erosional activity (Schuur et al., 2015) in part because long-term dynamics remain unclear.

Thermo-erosion may have intensified in recent decades in the Arctic. For example, the number and area of retrogressive thaw slumps (RTS) increased dramatically over the past ~50 years in northwestern Canada (e.g., Lantz and Kokelj, 2008; Lacelle et al., 2010). From 1997 to 2010, 80% of RTS initiation in the Noatak Basin of Alaska occurred during years with unusually warm spring temperatures and intense precipitation events (Balser et al., 2014). Thus, recent RTS activity may represent novel landscape disturbances attributable to anthropogenic climate change. However, remote-sensing analysis revealed an abundance of stabilized thermo-erosional features on Alaska’s North Slope (Bowden et al., 2012; Krieger, 2012). Although the age of these features is unknown, their abundance suggests that RTS activity occurred frequently in the past, possibly as a result of natural climate variability. Evaluating these alternative interpretations requires a long-term perspective.

We use lake-sediment records from the Alaskan Arctic (Fig. 3.1) to evaluate geochemical signatures of thermo-erosion and infer past RTS activity. Lake NE14 experienced recent thermo-erosion and has prominent RTS depressions in the watershed, whereas the Perch Lake watershed lacks such features. We conducted high-resolution elemental analysis of sediment cores from both sites spanning the past 6000 years using X-ray fluorescence (XRF). These elemental data are interpreted in conjunction with lake-water chemistry and stratigraphic variations in lithology, mineral composition, and Sr and C isotopes. These data provide new information on the patterns, processes, and drivers of thermo-erosion, with implications for Arctic landscape change in response to anthropogenic warming.

STUDY AREA

Lake NE14 (68°40′28″ N, 149°37′35″ W, 700 m a.s.l.) and Perch Lake (68°56′29″ N, 150° 29′58″ W, 400 m a.s.l.) are small, hydrologically open basins located ~50 km apart in the Arctic Foothills of Alaska (Fig. 3.1A), where 90–100% of the landscape is underlain by permafrost. Quaternary advances and retreats of Brooks Range mountain glaciers resulted in spatially complex surficial deposits in this region. Lake NE14 lies in till and moraine deposits of the Itkillik II glacial advance (25–11.5 ka) that contain relict glacial ice (Table 3.1). Perch Lake
lies in late Tertiary glacio-fluvial outwash deposits characterized by rocky soils with low ice content. Vegetation at both sites is tussock-sedge and dwarf-shrub tundra.

In 2005, a prominent RTS formed on the northeastern margin of NE14 and released fine-grained sediments into the lake (Fig. 3.1B), predominantly via an inlet channel that formed on the slump surface. Landsat imagery reveals that sediment flux from the slump dramatically changed lake-water color starting in summer 2005, and that this effect persisted in summer 2008 (Fig. B.1). By 2010 the slump headwall receded to its maximal position and the RTS was stabilized by early-successional vegetation. The shoreline of NE14 is characterized by several fan-shaped depressions (Fig. 3.1B), which likely represent older RTS events (Bowden et al., 2012). The 2005 RTS is situated within a larger depression. In contrast, Perch has no indication of thaw slumps in its watershed.

MATERIALS AND METHODS

During the summers of 2008–2011, we collected surface water samples from Lake NE14 (n = 114), the two major inlet streams (n = 46), and the channel that formed on the RTS surface (n = 23) (Table B.1). Surface water samples (n = 7) were also collected in 1988–2001, prior to the formation of the 2005 RTS. Lake-water electrical conductivity was measured with Orion meters. Water samples were filtered through pre-combusted Whatman GF/F filters prior to measurements of cation concentrations with a Perkin Elmer inductively coupled plasma spectrophotometer and total alkalinity with potentiometric titrations.

In summer 2011, we obtained stratigraphically overlapping sediment cores from the deepest area of a well-defined basin in each lake. Sediment chronologies span the past 6000 years based on AMS ¹⁴C dating of terrestrial macrofossils (Table B.2). Age-depth relationships were modeled with a cubic spline function through calibrated ¹⁴C ages and 2σ errors (Fig. B.3). Elemental composition of the cores was measured at contiguous 0.25-cm (NE14) and 1.0-cm (Perch) intervals with an Itrax XRF Core Scanner. The XRF peak intensities were measured as counts per second (cps) and normalized to coherent scatter. We also measured bulk density (g dry sediment cm⁻³) and loss-on-ignition (LOI at 950 °C) at contiguous 1-cm intervals for NE14 and at 4-cm intervals for Perch.

To help constrain interpretations of the XRF data, we selected sediment samples from NE14 for mineral and isotopic analyses. Mineral composition was determined with a ScintagR
XDS2000 X-Ray Diffractometer. δ^{13}C was measured before and after carbonate removal via acid-fumigation using a Thermo Fisher Delta V Advantage mass spectrometer. For $^{87}$Sr/$^{86}$Sr analysis, samples were fused with LiBO$_2$, dissolved in 3N HNO$_3$, purified with a Sr-spec resin column, and analyzed on a Nu Plasma HR MC-ICPMS spectrometer. The Sr isotope ratios were corrected based on NBS987 coral standards, and analytical precision is ± 0.00002.

To determine the ages of distinct Ca:Sr peaks in the NE14 record, we interpolated Ca:Sr samples to decadal resolution and used a robust loess smoother to remove the background trend. We combined nearby peaks and interpreted them as one RTS “episode”. To estimate the timing of RTS episodes, we adjusted the NE14 age-depth model by assuming instantaneous deposition of sediments in each episode (Fig. B.3D).

RESULTS AND INTERPRETATION

Surface water samples from NE14 prior to the 2005 RTS provide a baseline to assess changes in lake-water chemistry associated with thermo-erosion (Fig. 3.2). Prior to 2005, lake-water conductivity, alkalinity, and Ca concentration were 146 ± 7 µS cm$^{-1}$ (mean ± S.D.), 1476 ± 36 µeq L$^{-1}$, and 636 ± 79 µM, respectively. In 2008–2011, values were all significantly higher (t-test, $P$<0.001; Table B.1): 183 ± 25 µS cm$^{-1}$ for conductivity, 1959 ± 182 µeq L$^{-1}$ for alkalinity, and 867 ± 75 µM for Ca. From 2008 to 2011, water samples from the slump inlet also had significantly higher conductivity, alkalinity, and Ca, Mg, Na, and K concentrations compared with samples from inlets draining areas unaffected by the slump (t-test, $P$<0.0001). These data document the modern signature of thermo-erosion and aid in our interpretation of sediment XRF data.

The sediments from NE14 span the past 6000 years (0–189.5 cm; Figs. 2 and DR3C) and exhibit distinct stratigraphic variation. The sediments are predominantly gyttja with prominent light-gray beds comprised of clay-sized particles. These clay beds have well-defined bases and gradually transition to gyttja at the top, but show no evidence of turbidites associated with nearshore slides or underwater fans. The uppermost clay bed is 3–7 cm below the sediment-water interface and is interpreted as the 2005 RTS. The upper 3 cm of sediments are ~70% water, and thus represent ~0.9 cm of net sediment accumulation after influx from the 2005 RTS ceased (see DR3). Similar clay beds in the NE14 core span 24–28, 33–39, 49–63, 81–87.5, 100.5–122, and 138–158 cm, some of which contain alternating laminae of clay-rich sediments and gyttja.
XRF analysis shows that the clay beds correspond to peak values of Ca and elements associated with allochthonous sediments including Ti, K, Rb, Sr, and Zn (Fig. 3.3). Both %CaCO$_3$ and $\delta^{13}$C are generally higher in the clay than non-clay sediments (Fig. 3.4). Thus, the prominent Ca peaks in the NE14 core likely reflect carbonate-rich sediments associated with past RTS activity. This inference is supported by elevated Ca concentrations in lake-water samples after the 2005 RTS, and also in the inlet draining the slump than in the inlets draining the watershed.

In our study area, permafrost soils are characterized by high Ca:Sr ratios relative to the thawed active layer (Keller et al., 2007). This pattern reflects extensively weathered minerals in the active-layer, resulting in the dissolution of carbonate and preservation of silicates that contain abundant Sr, Ti, K, and Rb relative to Ca. In contrast, carbonate is abundant in permafrost soils, which are subject to weathering only when exposed by thaw or thermo-erosion. The influx of cations from permafrost soils to the lake may also have promoted authigenic calcite precipitation, further increasing the relative carbonate content of the sediments. Thus past RTS activity should result in elevated ratios of carbonate relative to silicate minerals in our sediment cores. Consistent with this expectation, the ratios of Ca:Sr and Ca:K in the NE14 sediments show dramatic peaks associated with clay beds (Fig. 3.4). Furthermore, the ratio of carbonate (calcite and dolomite) to feldspar and clay minerals is correlated with Ca:K and Ca:Sr ($r = 0.96$ and $0.93$, $P<0.0001$, $n = 15$).

In contrast to NE14, the Perch sediments of the past 6000 years (0–121 cm; Figs. 2 and DR3B) are predominately gyttja without prominent clay beds. The major elements that display pronounced downcore variation at NE14 exhibit little change at Perch. The most dramatic difference between the two sites is the generally low Ca content in the Perch sediments compared to distinct, high-amplitude Ca peaks corresponding to clay deposition from thermo-erosion at NE14. The %CaCO$_3$ of the Perch sediments is also low (2–5%) compared to NE14 (Fig. 3.4). Furthermore, the Ca:K and Ca:Sr ratios at Perch show little variation throughout the past 6000 years, suggesting no RTS activity.

Isotopic analyses of NE14 sediments lend additional support for our interpretations. Although no consistent $^{87}$Sr:$^{86}$Sr difference exists between the two sediment types, the average ratio is lower in clay ($0.724682 \pm 0.000601$, $n = 8$) than non-clay ($0.726284 \pm 0.000476$, $n = 8$) samples, suggesting different provenances. The $^{87}$Sr:$^{86}$Sr values are lower in limestone than non-limestone bedrock, and in deeper mineral soils compared to the active layer in this region (Keller
et al., 2007). Thus, lower $^{87}\text{Sr}:{^{86}\text{Sr}}$ ratios of clay samples at NE14 suggest the deposition of carbonate-rich material when permafrost thawed and thermo-erosion occurred. This allochthonous carbonate is enriched in $^{13}\text{C}$ relative to organic matter, as indicated by the $\delta^{13}\text{C}$ differences between bulk and acid-fumigated sediments (Fig. 3.4).

The geochemical signature of thermo-erosion at NE14 varies over the past 6000 years. The amplitude of Ca:Sr and Ca:K peaks displays a generally decreasing trend and the thickness of the clay beds varies through time (Figs. 2 and 3). These variations may reflect decreasing intensity of thermo-erosion or changing proximity of thaw slumps in the watershed to our core site. These alternative interpretations cannot be tested with our data. It is also difficult to determine how long each event lasted and if multiple small peaks within each clay bed represent a single or multiple RTS events. To help evaluate these alternatives, we obtained $^{14}\text{C}$ ages from terrestrial macrofossils (Table B.2) that bracket the thickest clay layer in the NE14 core (138–158 cm; Fig. B.4). The $2\sigma$ errors of these ages overlap: 3005–3380 cal BP (136.5 cm) and 3255–3466 cal BP (159.5 cm). One interpretation of this overlap is that the sediments deposited rapidly in a single event. This result is consistent with the rapid recovery of lake-water color (Fig. 3.1) and chemistry (Fig. 3.2) at NE14 following the 2005 RTS.

We use the adjusted NE14 age-depth model (Fig. B.3D) to determine the timing of past RTS episodes. Each episode may represent 1) several RTS events in rapid succession, 2) one RTS with multiple pulses of sediment influx over several years, or 3) one event followed by sediment reworking and redeposition. This analysis reveals 10 RTS episodes with $2\sigma$ age ranges of –56 to –47, 431–636, 658–841, 1116–1335, 1627–1817, 2039–2300, 2259–2551, 3104–3330, 3446–3726, and 5764–6075 cal BP (Fig. 3.4, right). These episodes represent a minimum estimate of past RTS events at NE14 over the past 6000 years.

**DISCUSSION**

The NE14 analyses provide one of the first continuous records of thermo-erosion spanning several millennia from the North American Arctic. Consistent with the lakeshore depressions suggestive of past RTS events, our sediment data indicate that the striking 2005 RTS is not unprecedented. No evidence exists that RTS are increasing in frequency or magnitude in the sediments of the 20th and 21st centuries at this site. In fact, the uppermost Ca:Sr peak is small and the clay bed thin compared to older RTS episodes (Fig. 3.4), suggesting that the most recent
RTS released less sediment than older events. A decline in the magnitude of RTS activity through time may have resulted from increasing thickness of organic soils during the Holocene, which insulate permafrost soils (e.g., Gaglioti et al., 2014). Alternative interpretations are possible, such as proximity of our core site to the location of RTS on the lakeshore. We cannot rigorously evaluate these alternative interpretations because we lack information on the deposition of eroded materials across the lake basin, which can be obtained with seismic profiling and sediment-core transects. Nonetheless, our results highlight repeated RTS episodes over the past 6000 years, and the importance of a long-term perspective to evaluating ongoing permafrost change.

The NE14 record exhibits temporal variation in the frequency of thermo-erosion (Fig. 3.4), and this variation may be linked to regional climate change. High summer temperatures can increase seasonal thaw depth in permafrost soils and thus favor thermo-erosion (e.g., Jorgenson et al., 2006). For example, cryostratigraphic analyses suggest that thermo-erosion increased in northwestern Canada in response to warm summers during the Holocene Thermal Maximum (Burn, 1997). Relatively little is known about the Holocene climate history of Alaska’s North Slope. However, within the range of $^{14}$C chronological uncertainty, the NE14 RTS record generally corresponds with temperature inferences from sediment chlorophyll content at Kurupa Lake (Boldt et al., 2015; Figs. 1A and 3). For example, several RTS episodes (3251, 2405, 2160, 1712, 1210, 755, and 532 cal BP) coincide with periods of elevated summer temperatures, whereas the interval of reduced thermo-erosion (5700–3720 cal BP) corresponds to lower temperatures. The broad consistency between the two records implies that summer temperatures played a major role in driving RTS occurrence at NE14 over the past 6000 years.

The oldest two episodes of thermo-erosion at NE14 (5425 and 3720 cal BP) occurred during a prolonged cool interval at Kurupa Lake, suggesting that factors other than summer temperature may contribute to thermo-erosion. Enhanced precipitation may facilitate RTS, both through initiating mass wasting that exposes ground ice and through removing slump sediments that sustain ice exposure (Lacelle et al., 2010). For example, moisture variability at millennial timescales strongly influenced thermo-erosion in Siberia and northern Alaska (Mann et al., 2010; Biskaborn et al., 2013). At NE14, increased effective moisture during the late Holocene (e.g., Clegg and Hu, 2010) may have contributed to the higher frequency of thermo-erosion over the past 4000 years than before (Fig. 3.4).
Thermo-erosion occurred repeatedly over the past 6000 years at NE14 but was absent at Perch, despite their proximity and the likelihood that the two sites experienced the same climate. Lake NE14 lies in morainal deposits with variable ice content, whereas Perch is located in glaciofluvial deposits with low ice content. Because higher ground-ice content leads to greater soil instability as permafrost thaws, differences in geomorphology and ground ice content may explain why the watershed of NE14 is more prone to RTS activity than the Perch area. However, many lakes located in the same terrain as NE14 do not have visible RTS features in their watersheds, suggesting no recent thermo-erosion. One plausible explanation is that positive feedbacks facilitated RTS reactivation at NE14, as evidenced by the fact that the 2005 RTS is nested within a larger depression. For example, removal of surface soils reduces insulation in summer and enhances snow accumulation in thaw slumps, increasing ground heat flux and promoting reactivation (e.g., Osterkamp et al., 2009). Changes to the ground thermal regime can result in talik formation, which promotes further thaw-induced subsidence and RTS growth (Kokelj et al., 2009). Thus, once initiated, thermo-erosion may have heightened sensitivity to warming with a tendency to occur repeatedly in the same areas.

Permafrost thaw and thermo-erosion are anticipated to become more prevalent if the trend of Arctic warming continues (Jorgenson et al., 2006; Schuur et al., 2015). The potential for positive feedbacks leading to repeated thermokarst failures, as suggested by the NE14 record, highlights the instability of Arctic landscapes in a changing climate. Increased thermo-erosion may catalyze many other changes, such as shrub encroachment in tundra regions (Schuur et al., 2007; Osterkamp et al., 2009). Increased shrub density alters albedo and facilitates burning, which may further promote thermo-erosion via changes in the ground thermal regime (Hu et al., 2015). Furthermore, thermo-erosion can release vast amounts of labile carbon in ground ice deposits (Schreiner et al., 2014), which can elevate atmospheric CO₂ concentration and accelerate climate warming.

This study offers insight into the temporal variability and drivers of thermo-erosion at a single site, and the response of Arctic landscapes to climate change will likely be heterogeneous. RTS formation results from the wastage of buried ice in ice-rich landscapes, including recently deglaciated areas and older surfaces underlain by ice wedges (Swanson, 2014). However, ongoing processes such as paludification may insulate ice-rich soils against climate-driven thaw (Gaglioti et al., 2014). Thus, spatial variability in ground ice, glacial history, and landscape
evolution will impact the response of Arctic landscapes to climate-driven permafrost thaw. Understanding the patterns and drivers of thermo-erosion and its linkages to other surficial processes requires additional research to capture this variability. Our study offers one of the first millennial-scale records contributing to this understanding and provides a long-term context for evaluating ongoing Arctic change.

**TABLES**

**Table 3.1:** Site characteristics for NE14 and Perch Lake. Land-cover classification based on the Circumpolar Arctic Vegetation Map (CAVM; Walker et al., 2005). Terrain, thermokarst features, and ice content based on Jorgenson et al. (2008). Surficial geology from NE14 and Perch based on Hamilton (1978) and (1979), respectively. Ages of the surficial deposits from Hamilton (1986). Bedrock units from Beikman (1980).

<table>
<thead>
<tr>
<th>Site</th>
<th>NE14</th>
<th>Perch</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Lake</strong>*</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Longitude</td>
<td>149°37'35&quot; W</td>
<td>150°29'58&quot; W</td>
</tr>
<tr>
<td>Elevation (m a.s.l.)</td>
<td>700</td>
<td>400</td>
</tr>
<tr>
<td>Surface area (ha)</td>
<td>25.4</td>
<td>14</td>
</tr>
<tr>
<td>Perimeter Length (km)</td>
<td>2.4</td>
<td>1.6</td>
</tr>
<tr>
<td>Maximum water depth (m)</td>
<td>20</td>
<td>12.6</td>
</tr>
<tr>
<td>Water depth, core site (m)</td>
<td>13.2</td>
<td>12.6</td>
</tr>
<tr>
<td><strong>Vegetation</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CAVM land-cover class</td>
<td>Tussock-sedge, dwarf-shrub tundra</td>
<td>Tussock-sedge, dwarf-shrub tundra</td>
</tr>
<tr>
<td><strong>Permafrost</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Terrain</td>
<td>Modified moraine</td>
<td>Glacio-fluvial</td>
</tr>
<tr>
<td>Thermokarst features</td>
<td>Thermokarst lakes and slumps</td>
<td>Troughs and pits</td>
</tr>
<tr>
<td>Ice content</td>
<td>Variable</td>
<td>Low</td>
</tr>
<tr>
<td><strong>Surficial Geology</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Unit</td>
<td>Itkillik II Glacial Drift**</td>
<td>Gunshot Mountain Glacial Drift</td>
</tr>
<tr>
<td>Age</td>
<td>11.5-25 ka</td>
<td>Late Tertiary</td>
</tr>
<tr>
<td>Description</td>
<td>Poorly-sorted till and stratified ice-contact deposits</td>
<td>Highly eroded glacial deposits overlain by &gt;2 m organic silts</td>
</tr>
<tr>
<td><strong>Bedrock Geology</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bedrock Age</td>
<td>upper Cretaceous</td>
<td>lower Cretaceous</td>
</tr>
<tr>
<td>Bedrock Type</td>
<td>continental sedimentary deposits</td>
<td>continental sedimentary deposits</td>
</tr>
</tbody>
</table>

*Lake NE14 and Perch Lake (or Perched Lake) are Arctic Long-Term Ecological Research (LTER) sites.
**NE14 lies in the Itkillik Valley Lobe deposits (id2), which are described as extensive ice stagnation deposits, with irregular kettles (Hamilton, 2002). The lakes in this area are defined as low Arctic kettle lakes (Hobbie and Kling, 2014).
**Figure 3.1:** A) Locations of Lake NE14, Perch Lake, and Kurupa Lake (Boldt et al., 2015). B) Aerial photograph of NE14 (18 July 2007) with core site (dot), RTS (arrow), and blue lake water from sediment influx. C) NE14 with stabilized slump headwall and clear lake water (Google™ Earth; 17 July 2009)
Figure 3.2: Chemical measurements of water samples from Lake NE14 and its inlets (mean ± S.D., standard deviation). Pre-event lake-water samples were obtained CE 1988-2001, and post-event lake and inlet samples were obtained CE 2008-2011. Units are μS cm\(^{-1}\) for conductivity (Cond), μeq L\(^{-1}\) for alkalinity (Alk), and μM for cations.
Figure 3.3: Elemental composition of NE14 (black) and Perch (gray) sediments (see Fig. B.4). Normalized counts per second are raw counts divided by coherent scatter.
Figure 3.4: Evidence for RTS episodes at NE14: BD (bulk density), %CaCO₃ (950 °C LOI), Ca:K and Ca:Sr (black and gray for NE14 and Perch, respectively), δ¹³C (black and white for bulk and acid-fumigated samples, respectively), Carb:Non-Carb (carbonate:feldspar+clay minerals), ⁸⁷Sr:⁸⁶Sr (reversed x-axis), and Ca:Sr Peaks (background removed). Horizontal gray bars denote RTS episodes. JJA (June-August) temperature anomalies (Boldt et al., 2015), with chronology shifted older by 150 years (within ¹⁴C chronological uncertainty); this offset is arbitrary and we do not know why it results in a better correlation. The right y-axis shows age and 2σ range of RTS episodes assuming instantaneous deposition.
REFERENCES

Balser, A.W., Jones, J.B., and Gens, R., 2014, Timing of retrogressive thaw slump initiation in
the Noatak Basin, northwest Alaska, USA: Journal of Geophysical Research, v. 119,
p. 1106–1120.

500 000, 2 sheets.

Biskaborn, B.K., Herzschuh, U., Bolshiyov, D.Y., Schwamborn, G., and Diekmann, B., 2013,
Thermokarst processes and depositional events in a tundra lake, northeastern Siberia:

reconstruction from sedimentary chlorophyll content, with treatment of age uncertainties,
Kurupa Lake, Arctic Alaska: The Holocene, v. 25, p. 641–650,

Bowden, W.B., et al., 2012, An integrated assessment of the influences of upland thermal-
erosional features on landscape structure and function in the foothills of the Brooks
Range, Alaska, in Hinkel, K.M., ed., Tenth International Conference on Permafrost,
Volume 1: International Contributions, Salekhard, Russia, The Northern Publisher, p. 61–
66.

Burn, C.R., 1997, Cryostratigraphy, paleogeography, and climate change during the early
Holocene warm interval, western Arctic coast, Canada: Canadian Journal of Earth

south-central Brooks Range, Alaska: Quaternary Science Reviews, v. 29, p. 928–939,

Cory, R.M., Crump, B.C., Dobkowski, J.A., and Kling, G.W., 2013, Surface exposure to sunlight
stimulates CO$_2$ release from permafrost soil carbon in the Arctic: Proceedings of the

Radiocarbon age-offsets in an arctic lake reveal the long-term response of permafrost

Hamilton, T.D., 2002, Glacial Geology of the Toolik Lake and the Upper Kuparuk River Region:
Biological Papers of the University of Alaska, no. 26, 25 p.

Society, 49 p.

Geological Survey Miscellaneous Field Studies Map 1121, scale 1:250,000, 1 sheet.

Hamilton, T.D., 1978, Surficial geologic map of the Phillip Smith Mountains quadrangle,
1:250 000, 1 sheet.


CHAPTER 4: LATE-HOLOCENE INTERACTIONS BETWEEN CLIMATE, FIRE, AND THERMO-EROSION: IMPLICATIONS FOR NOVEL TUNDRA DISTURBANCE REGIMES

ABSTRACT
Amplified warming in the Arctic may facilitate novel disturbance regimes in tundra ecoregions, as suggested by recent increases in the rate and extent of thermo-erosion and the occurrence of record-setting fires in some tundra areas. Both thermo-erosion and wildfire can exacerbate warming via release of large carbon stocks in permafrost soils. In addition, these disturbance regimes may interact, which can lead to complex ecosystem feedbacks. However, the interaction of tundra disturbance regimes is difficult to assess because of spatial variability of ground-ice deposits and associated thermo-erosion, rarity of modern tundra fires, and short observational records. We overcome these limitations by using geochemical and charcoal records from the late-Holocene sediments of Loon Lake in the Noatak National Preserve, Alaska, to identify thermo-erosional episodes and fire events over the past 3000 years. These data are interpreted in the context of soil geochemistry from modern retrogressive thaw slumps (RTS). Magnetic susceptibility (MS), and the ratios of Ca:Sr and Ca:K increased with depth in modern RTS soils and were higher on recently-exposed than older slump surfaces. Peaks in bulk density, CaCO$_3$, MS, Ca:K, and Ca:Sr values (summarized with PCA) in the Loon Lake sediments suggest a minimum of 18 RTS events in the watershed over the past 3000 years. Using macroscopic charcoal, we identified 22 fire events at Loon Lake and compared fires to the thermo-erosional record using superposed epoch analysis. Temporal variability in our records suggests climate-driven responses of both thermo-erosion and fire disturbance regimes, with lower-magnitude PC axis 1 peaks and fewer charcoal-inferred fire events during the cool Little Ice Age than the warm Medieval Climate Anomaly. In addition, the MS, Ca:K, Ca:Sr, and PC axis 1 values show a significant response (>90% CI) to catchment fires within 20-30 years of fire occurrence, suggesting soil warming following fires may facilitate thermo-erosion on decadal timescales. Our results are the first to assess the long-term relationship between fire and thermo-erosion in Arctic tundra, and highlight the potential for interacting tundra disturbance regimes in response to ongoing climate change.
INTRODUCTION

Anthropogenic warming is amplified in the Arctic (Serezze and Barry, 2011), and has resulted in profound environmental changes that may further enhance global warming via positive feedbacks (Chapin et al., 2005; Post et al., 2009). Arctic ecosystems contain large carbon stocks in frozen soils (Tarnocai et al., 2009; Hugelius et al., 2014) and ground-ice deposits (Fritz et al., 2015). Pulse disturbance processes such as wildfire and thermo-erosion (i.e., the subsidence and lateral transport of ice-rich sediments) are key pathways for the rapid release of soil carbon in permafrost areas (Schuur et al., 2008; Grosse et al., 2011). Wildfires release near-surface carbon stocks via combustion of vegetation and soil biomass (Mack et al., 2011). Thermo-erosion can facilitate the transport of soil carbon to aquatic systems (Vonk and Gustafsson, 2013; Cory et al., 2014), as well as expose ancient carbon-rich soils to microbial decomposition (Cory et al., 2013). These disturbances can impact surface hydrology (Grosse et al., 2011; Schuur et al., 2015), mobilize vast glacial sediment stores (Kokelj et al., 2017), modify vegetation composition (Sturm et al., 2005; Schuur et al., 2007; Osterkamp et al., 2009; Walker et al., 2009), and alter aquatic community structure (Bowden et al., 2008; Mesquita et al., 2010; Thompson et al., 2012; Thienpoint et al., 2013). Thus, feedbacks between climate, disturbance, and ecosystem processes will likely play an important role in the complex response of Arctic ecosystems to climate change (Bowden, 2010).

Ongoing climate change may impact the thermo-erosional regime, as suggested by recent increases in the rate and extent of thermo-erosion in several ice-rich regions of the Arctic (e.g. Jorgenson et al., 2006; Bowden et al., 2008; Lantz et al., 2008; Lacelle et al., 2010) and evidence for enhanced thermo-erosional activity in the North American Arctic coincident with periods of elevated summer temperatures during the Holocene (Burn, 1997; Lacelle et al., 2004; Mann et al., 2010; Chipman et al., 2016). However, the sensitivity of permafrost terrain to climate warming is dependent on many factors that are spatially heterogeneous, including soil texture and geomorphology, the distribution of ground ice deposits, and landscape history (Mann et al., 2010; Biskaborn et al., 2013; Gagliotti et al., 2014; Segal et al., 2016; Kokelj et al., 2017). Climate warming may also indirectly impact permafrost soils by influencing precipitation patterns and the timing of snowmelt, which are important controls on thermo-erosion in upland tundra areas with ice-rich soils (Lacelle et al., 2010; Balser et al., 2014; Kokelj et al., 2015; Segal et al., 2016). Although these recent studies have improved our understanding of the
climate and landscape drivers of permafrost thaw in ice-rich terrain, we still know little about spatiotemporal patterns of thermo-erosion.

Recent record-setting fire activity in tundra ecoregions of Alaska (Jones et al., 2009; Rocha et al., 2012) suggest that Arctic warming can facilitate burning in areas where cold climates and limited biomass have historically inhibited fire activity (Hu et al., 2015). Because the modern occurrence of tundra fires is strongly linked to climate, the frequency and severity of fires throughout Alaskan tundra regions are expected to increase dramatically in the future (Hu et al., 2015; Young et al. 2016), which may result in novel fire regimes in areas that have not burned for thousands of years (Hu et al., 2010; Chipman et al., 2015). Wildfire may be a key climate-driven mechanism for inducing catastrophic permafrost thaw, because the removal and/or reduction of vegetation and organic soils following fire results in loss of thermal insulation, reduced surface albedo, and reduced heat loss from evapotranspiration (Rocha et al., 2012). These changes may in turn promote higher soil temperatures and active-layer thickening, leading to rapid thermo-erosion and organic carbon losses in ice-rich terrain (e.g., Liljedahl et al., 2007; Lacelle et al., 2010; Jones et al., 2015). Thus, the interaction between fire and thermo-erosion, combined with the direct impacts of climate warming, may have profound consequences for the fate of Arctic soil carbon. However, we currently know little about the long-term relationship between changing fire and thermo-erosional regimes in tundra ecosystems.

Understanding the interaction between fire and thermo-erosional activity is critical for anticipating the consequences of climate change in the Arctic. Fires have been rare in most tundra ecoregions (Chipman et al., 2015), and thus it is difficult to examine the role of fire on thermo-erosional activity. In tundra areas where fires are common, examining this relationship is complicated by spatial heterogeneity in ground-ice deposits, which is not well quantified at the landscape scale where these interactions occur. Moreover, observations of both fire and thermo-erosional activity are short (~60 years), which may obscure long-term feedbacks between these disturbance regimes. These limitations can be circumvented by using a paleoecological approach to examine the relationship between fire and thermo-erosional activity at the same site through time. In this study, we use a combination of magnetic susceptibility and X-ray fluorescence to reconstruct past retrogressive thaw slump (RTS) episodes at Loon Lake in the Noatak National Preserve. The interpretations of our sedimentary and geochemical data are aided by soil profiles
from RTS features at Loon Lake and nearby Bonus Lake. We use macroscopic charcoal to
reconstruct local fire events at Loon Lake, and assess the interaction between fire and thermo-
erosion over the past 3000 years using superposed epoch analysis (SEA). To our knowledge, this
is the first study to explicitly test the long-term relationship between fire and thermo-erosion in
the Arctic tundra. We chose the Noatak National Preserve for this study because thermo-
erosional features are present throughout the region (Swanson and Hill, 2010; Balser et al.,
2014) and fire has been a common occurrence on the landscape for at least the past 6000 years
(Higuera et al., 2011a). The Noatak is also one of the most flammable tundra regions in Alaska
at present (Rocha et al., 2012) and may thus be an analogue for other tundra ecoregions in the
future as Arctic temperatures continue to increase.

STUDY AREA

The Noatak National Preserve (Fig. 4.1A) is a low-Arctic tundra ecoregion in
northwestern Alaska. In nearby Kotzebue (~115 km south of our study sites), mean June-August
and December-February temperatures are 10.1 and -18.9 °C, respectively (WRCC 1949-2015
observations). Mean annual precipitation is 252 mm, with 52% falling between July and
September, and mean annual snowfall is 139 cm, with 95% occurring from October to April.
Vegetation in the Noatak ecoregion is predominantly low-shrub and herbaceous tundra, with
sparse alpine vegetation at high elevations and isolated pockets of tall-shrub and boreal
vegetation in the western portion of the watershed (Swanson, 2015). The Noatak ecoregion has a
modern fire-return interval of ~425 years (Rocha et al., 2012), and the fire-regime in the western
Noatak has been active for at least the past 6000 years, as suggested by charcoal-based
reconstructions showing individual fire-return intervals of 30-840 years (Higuera et al., 2011a).

The Noatak ecoregion is underlain by continuous permafrost (Brown et al., 1997).
Glacial and periglacial deposits dominate the surficial geology, reflecting multiple Pleistocene
 glaciations that shaped the landscape (Hamilton, 2010). Moraine, drift, and glaciolacustrine
deposits are common (Jorgensen et al., 2008), some of which contain ice wedges and relict
 glacial ice (Swanson and Hill, 2010). Permafrost disturbance includes RTS, active-layer
detachment slides, and thermo-erosional gulleys (Swanson and Hill, 2010; Balser et al., 2014).
RTS features tend to form in Late Pleistocene glacial till deposits that have deep relict ice, and
are often found on northwest-facing slopes along rivers and lakes (Fig. 4.1A; Swanson, 2014).
Recent analyses suggest that these modern RTS features are sensitive to weather, particularly the length of the spring season and the intensity of precipitation events (Balser et al., 2014).

Our study sites are located in the western portion of the Noatak National Preserve (Fig. 4.1A). Both Loon Lake (67°55'45.37"N, 161°58'0.95"W, 90 m a.s.l.; Fig. 4.1B) and Bonus Lake (67°56'42.40"N, 162°21'18.94"W, 150 m a.s.l.; Fig. 4.1C) lie in isolated, coarse-grained deposits of late-Pleistocene to early-Holocene colluvium (Hamilton, 2010) and have prominent RTS features within their watersheds. The vegetation cover in the western Noatak is a combination of birch-Ericaceous shrub tundra, willow shrub tundra, and tussock-shrub tundra, which include Betula, Alnus, Salix, Poaceae, and Cyperaceae, as well as Picea glauca along river drainages (Jorgensen et al., 2009). In the summer of 2012, the most distinct RTS at Loon Lake (NOAT0788) was vegetated by early-successional graminoids, clumps of Salix near the shoreline, and Betula and Ericaceae on remnant pre-slump surfaces. The older slump surfaces (to the northeast of the most recent RTS) were densely vegetated, with the smaller RTS (NOAT0798) dominated by Alnus, Salix, and Equisetum, and the larger RTS (LOON01) containing abundant Betula and Ericaceae. The Bonus Lake RTS (NOAT0750) was sparsely vegetated by early-successional graminoids and stands of Alder and Salix.

METHODS

Geochemical Analyses of Soils and Lake Sediments

In the summer of 2012, we obtained samples from the exposed headwall soils of the most distinct RTS at both Bonus and Loon lakes (Fig. 4.1). We also collected samples from exposed mineral soils at the top and base of the slump floors, and from vegetated slump floors. We classify the vegetated soil samples as follows: sparsely vegetated surfaces characterized by graminoids (early-successional), vegetation cover dominated by grasses, Salix, and/or Alnus (mid-successional), and dense shrub cover with little to no soil exposure (late-successional).

For paleoecological analyses, we obtained sediment cores that were stratigraphically overlapping from the deepest area (12.2 m) of a well-defined basin in Loon Lake using a modified Livingstone piston corer. To assure an intact sediment-water interface, we used a polycarbonate tube fitted with a piston to retrieve the uppermost sediments. The unconsolidated sediments at the top of the core (0-6.5 cm) were subsampled into whirl-pack bags in the field,
and all remaining core segments were split lengthwise and correlated based on visual stratigraphy to assure a continuous core sequence in the laboratory (Fig. C.1).

Five terrestrial macrofossils were obtained from the lake-sediment core for AMS radiocarbon dating (Table C.1). Macrofossils were treated with an acid-base-acid protocol (Oswald et al., 2005), and submitted to Lawrence Livermore National Laboratory. Radiocarbon ages were calibrated to years before present (before 1950) using the Intcal13 dataset (Reimer et al., 2013) in Calib v7.0.4 (Stuiver and Reimer, 1993). The uppermost 15 cm of sediments were processed for \(^{210}\)Pb analysis following Eakins and Morrison (1978). \(^{210}\)Pb activity was measured on an Octête Plus alpha spectrometer, and sample ages were estimated following Binford (1990). An age-depth model was developed from \(^{210}\)Pb and calibrated \(^{14}\)C ages (± 2 S.D.) using a cubic spline function in CLAM v2.2 (Blaauw, 2010) with 10,000 bootstrapped iterations.

The elemental composition of the soil samples and the lake-sediment core was measured with an Itrax XRF Core Scanner at the University of Minnesota, Duluth. The soil samples and unconsolidated uppermost-6.5 cm sediments were packed in 16.4-cm\(^3\) sample holders for XRF scanning, and the remainder of the sediment core from Loon Lake was scanned at contiguous 0.5-cm intervals. The XRF peak intensity for each element was measured as counts per second (cps) and normalized to coherent scatter. For the lake-sediment core, loss-on-ignition was measured at contiguous 0.5-cm intervals at 550 °C for 2 hours and 950°C for 4 hours to calculate organic matter (%OM) and carbonate content \([\%\text{CaCO}_3 = \text{LOI} \% \times (100/44)]\), respectively. Magnetic susceptibility was measured on individual samples at contiguous 0.25-cm intervals using a Bartington Instruments MS2C meter. The mean temporal resolution of the XRF and MS records over the past 3000 years are 8 and 4 yr sample\(^{-1}\), respectively. To identify past thermo-erosional episodes, we used Principal Components Analysis (PCA) to summarize the MS, Ca:K, Ca:Sr, % CaCO\(_3\), and bulk density values and interpreted positive PC1 values as intervals of past episodes of RTS activity.

**Lake Sediment Charcoal Analyses**

For charcoal analyses, lake-sediment samples of 1-3 cm\(^3\) each were placed in a 20-ml solution of 5.25% bleach and 10% sodium metaphosphate for approximately 24 h and washed through a 120-μm sieve. Charcoal particles >120 μm were enumerated at 10-40X magnification and used to calculate charcoal accumulation rate (CHAR; pieces cm\(^{-2}\) yr\(^{-1}\)).
CharAnalysis v 1.1 (Higuera et al., 2009) to infer the timing of local fires (i.e., within 0.5-1 km of the lake; Higuera et al., 2007). The mean temporal resolution of the charcoal record over the past 3000 years is 4 yr sample\(^{-1}\). Prior to statistical analyses, CHAR values were interpolated to decadal resolution, and low-frequency (i.e., background) CHAR was estimated with a 750-yr robust loess regression and subtracted from the interpolated series. A Gaussian mixture model was used to separate the detrended CHAR series into noise and peak distributions, with the threshold of the peak series defined by the 99\(^{th}\) percentile of the noise distribution. Peak values based on charcoal counts that differed statistically (\(P\)-value <0.1) from the non-peak values of the previous 150 years were interpreted as local fire events. In addition to detecting local fires, we identify high-CHAR fires by removing fires with CHAR peak magnitudes in the first quartile range of all peak values (< 0.25 particles cm\(^{-2}\) peak\(^{-1}\), n=5), assuming that fires with low charcoal particle counts were either small or occurred outside the catchment (Whitlock et al., 2006; Dunnette et al., 2014). We estimated the age of all fire events as the age of the sample with the largest charcoal particle count within each decadally interpolated CHAR peak.

To estimate regional trends in biomass burning over the past 3000 years, we created a composite CHAR record by combining the charcoal record from Loon Lake with four additional lake-sediment charcoal records from the western Noatak (Higuera et al., 2011a; Fig. 4.1). The five charcoal records were composited following the method of Kelly et al. (2013), which standardizes the data based on the assumption that CHAR values fit a zero-inflated log normal (ZIL) distribution, and uses kernel-weighted local likelihood estimations to pool the records. This method is appropriate for the Noatak charcoal records because they comprise discrete particle counts and contain a high number of samples with counts equal to zero. We applied a Gaussian kernel with a bandwidth of 25 years (i.e., centennial-scale variability) to the standardized CHAR values at annual target points, and defined the composite CHAR value at each point as the mean of the best fit ZIL distribution. Parametric bootstrapped sampling of the fitted distribution was used to derive 95\% confidence intervals.

**Analysis of Fire and Thermo-erosion Relations**

Our charcoal and geochemical proxies are from the same sediment sequence, permitting comparison of fire and thermo-erosion events without complications related to chronological errors. To assess the relationship between fire events and thermo-erosional activity, we used
Superposed Epoch Analysis (SEA), which is a data composting technique that examines the response of time series data to the timing of discrete events at multiple lagged windows (i.e., pre- and post-event), and has been used to decipher ecosystem responses to CHAR-inferred fires in paleorecords (e.g., Blarquez and Caraillet, 2010; Dunnette et al., 2014; Leys et al., 2016). Using SEA, we examined the response of decadally interpolated MS, Ca:Sr, Ca:K, and PC1 records to the timing of all CHAR peaks (n=22) and high-CHAR peaks (n=17). The background trend was removed from time series data using a robust loess smoother (750-yr window), and the influence of the charcoal peaks on the residual response variables was measured at 10-yr time steps from 50 years before to 100 years after fire events. The 90, 95, and 99% confidence intervals were obtained by extracting the appropriate percentiles from 10,000 randomized composites (i.e., Monte Carlo method), accounting for temporal autocorrelation with randomized block bootstrapping (bin size = 5 samples). The SEA analysis was performed in Matlab using custom scripts from Dunnette et al. (2014).

RESULTS

Soils obtained from the RTS headwalls and slump floors show an increase in Ca:K and Ca:Sr with depth (Fig. 4.2). Magnetic susceptibility (MS) values, which reflect the mineral content, are higher in samples from the base than the top of the slump floors. Furthermore, the MS values from exposed soils on the slump floor at Loon Lake are an order of magnitude higher than those of the headwall soils. Younger thermo-erosional features with early-successional vegetation have higher MS and Ca:K values than older, late-successional surfaces at both sites, although Ca:Sr values do not differ significantly between surfaces at Loon Lake.

We retrieved a 278.5-cm sediment core from Loon Lake (Fig. 4.3 and C.1). We focus our analyses on sediments from 0-195.5 cm, which span the past 3000 years and have relatively robust chronological control (Fig. 4.4). These sediments are predominantly alternating beds of clay and gyttja, with finely laminated sediments (laminae <1 cm in thickness) from 66.5-90 cm and 116.5-120.5 cm. Generally, clays, clay-rich gyttja, and finely laminated sediments have higher abundances of elements associated with allochthonous inputs (i.e., Ca, Sr, Si, K, Ti, Rb, and Fe) compared to gyttja sediments (Fig. 4.3). The percentage of CaCO₃ and values of MS, Ca:Sr, and Ca:K are also higher for clay-rich than gyttja sediments and generally exhibit the same temporal pattern (Fig. 4.5). We assume that concurrent and distinct peaks in these values,
which are summarized by PCA, reflect the influx of mineral-rich soils from RTS activity. PC axis 1 explains 73.5% of the total variation in the dataset, and decadally-interpolated positive PC1 values are used to identify sediments associated with watershed thermo-erosion. This analysis reveals a minimum of 18 episodes of past RTS activity at Loon Lake (Fig. 4.5, grey bars) that correspond to clay-rich sediments at 13-14, 21.5-23, 27-28.5, 33.5-37.5, 39-45, 48-55, 56-57, 60-66, 67.5-75, 76.5-89, 99-102, 117-119, 127-135, 137-138, 149-150, 151.5-155.5, 158-161, and 186-194 cm.

Charcoal peak analysis reveals 22 local fire events (i.e., within 1 km) at Loon Lake over the past 3000 years (Fig. 4.6). The most recent CHAR peak spans -44 to -54 cal BP, corresponding to the fire that occurred in the watershed C.E. 1999 (Fig. 4.1; note that the most recent fire occurred in 2012, after core retrieval). The mean (95% C.I.) of the fire-return intervals during the past 3000 years is 140 (97-196) years. Five CHAR peaks at 877, 1253, 1673, 1932, and 2113 cal BP were characterized by low peak magnitudes (<0.25 particles cm\(^{-2}\) peak\(^{-1}\); Fig. 4.6, blue dots), and the 17 remaining peaks are defined as high-CHAR fires. The composite CHAR record shows broad intervals of high values 700-900, 1200-1400, and 2600-2900 cal BP, and overall moderate but variable values 1600-2300 cal BP (Fig. 4.6). Composite CHAR values are close to zero from 100 to 500 cal BP and display an increasing trend from 100 cal BP to present.

The SEA analysis provides an assessment of the linkages of RTS activity with individual CHAR-inferred fire events at Loon Lake. Twelve of the 22 decades with significant CHAR peaks either overlap or occur within three decades of an RTS episode (Fig 4.6), and the SEA shows that residual MS, Ca:K, Ca:Sr, and PC1 values increase within one to three decades after local fire events (Fig. 4.7, left column). However, only the 20-year lagged response of the MS values to local fire events is significant in the SEA. In contrast, all of the response records show a significant increase in residual values (90% C.I. or higher) 20 to 30 years after high-CHAR fires (Fig. 4.7, right column).

**DISCUSSION**

*Inferring thermo-erosion from soil and lake-sediment data*

Our soil data from the RTS features at Bonus and Loon lakes document the sedimentary and geochemical signals of modern thermo-erosion. At these sites, MS and Ca:K are higher in
the soils of younger than older slump surfaces (Fig. 4.2), reflecting greater mineral content and less weathering in recently exposed surfaces compared to older, stabilized RTS features. In addition, MS, Ca:K, and Ca: Sr values increase with soil depth (Fig. 4.2). Deeper soils have abundant Ca relative to K, Sr, Ti, and Rb because of more extensive weathering in the active layer of permafrost soils near the surface, which would have preferentially removed carbonate relative to silicate minerals (Keller et al., 2007). We use these soil data to help decipher the lake-sediment record from Loon Lake and infer thermo-erosional activity over the past three millennia.

Permafrost thaw and thermo-erosion can release abundant mineral particles and cation-rich meltwater from previously frozen soils to nearby lakes and streams, resulting in increased mineral content in lake sediments and cation abundance in lake water (Kokelj et al., 2005; 2013; Keller et al., 2010; Chipman et al., 2016). At Loon Lake, the clay-rich sediments with elevated CaCO$_3$, Ca: Sr, and Ca:K values (Fig. 4.5) likely reflect the influx of carbonate-rich materials from permafrost soils during RTS formation and/or precipitation of authigenic calcite in response to increased concentrations of Ca$^+$ in the lake water. These sediments also exhibit higher MS values than non-clay sediments, suggesting erosion of lithic material from the catchment (Fig. 4.5). These lithological and geochemical variations suggest a minimum of 18 episodes of thermo-erosion at Loon Lake over the past 3000 years, at approximately 160-180, 300-330, 400-420, 510-580, 600-710, 750-870, 890-910, 950-1060, 1080-1200, 1220-1420, 1570-1620, 1830-1860, 1960-2060, 2080-2090, 2220-2230, 2250-2300, 2330-2370, and 2800-2960 cal BP (Fig. 4.5, grey bars). Each episode inferred from our sediment record may represent an instantaneous event, an event with sediment reworking following deposition, or an interval with several pulses of sediment influx following initial RTS formation. RTS features can stabilize within several years (Jones et al., 2015; Chipman et al., 2016) or remain active over several decades depending on ground-ice content and duration of ice exposure (Burn and Friele 1989; Kokelj et al., 2005). The span of the RTS episodes at Loon Lake range between a single decade to 200 years. It is unlikely that these episodes represent instantaneous deposition, given the variability in both the values of the geochemical indicators (Fig. 4.5) and charcoal particle counts (Fig. 4.6; see next section) within RTS sediments. Furthermore, the longest intervals of RTS activity are characterized by laminated clays with thin bands of gyttja and/or coarse sediments (Fig. 4.3), suggesting either several RTS events in rapid succession and/or reactivation of the same RTS
features over long time periods. Thus, the 18 thermo-erosional episodes identified from the Loon Lake record likely represent a minimum estimate of RTS events over the past 3000 years.

The thermo-erosional record from Loon Lake may be linked to climate change over the past three millennia. The thickness of the clay-rich sediments derived from RTS episodes, as well as the amplitude of peaks in the geochemical records, probably reflect the severity of past thermo-erosional activity. Over the past 2000 years, high MS, CaCO3, Ca:K, Ca: Sr, and PC1 peaks associated with relatively thick (7.0-12.5 cm) clay bands occurred during the Medieval Climate Anomaly (MCA, ~750-1200 cal BP) when the regional climate was relatively warm (Hu et al., 2001; D’Arrigo et al., 2006; Wiles et al., 2008; Boldt et al., 2015) (Fig. 4.5). Conversely, small peaks and relatively thin clay bands (1.0-6.0 cm thickness) coincided with cooler temperatures and glacial advances during the Little Ice Age (LIA, ~70-770 cal BP; Barclay et al., 2009). The highest MS, Ca:K, and Ca: Sr values of the record occur between 2000 and 3000 cal BP, which broadly overlaps with an interval of relatively warm summer temperatures in the northern Brooks Range (Boldt et al., 2015). Within this interval, no thermo-erosional events are identified during periods of mountain glacier advance in northern Alaska 2700-2600 cal BP (Calkin, 1988; Badding et al., 2013). However, discrepancies exist in the inferences of RTS episodes at Loon Lake and regional climate. For example, large peaks occurred near the end of the 2200-2000 and 3000-2900 cal BP glacial advances in the Brooks Range (Badding et al., 2013; Solomina et al., 2015; Kaufman et al., 2016). In addition, two thermo-erosional episodes overlap with the first millennial cooling in Alaska (1350-1750 cal BP; Wiles et al., 2008), although the peak magnitudes from the episode at 1570-1620 cal BP are relatively small. These discrepancies may be related to spatial heterogeneity in temperature changes and/or other factors, such as moisture variability and fire, controlling thermo-erosional activity at Loon Lake.

We compare the Loon Lake thermo-erosional record to RTS activity inferred from geochemical and lithological indicators at Lake NE14, located in the Arctic Foothills, north of the Brooks Range (Fig. 4.1, inset; Chipman et al., 2016). The sample resolution of the XRF records at Lake NE14 (6 yr sample⁻¹) is comparable to that at Loon Lake (8 yr sample⁻¹), and records at both sites were decadally interpolated to identify past thermo-erosional episodes. Comparison of the Lake NE14 record to a climate reconstruction from the northern Brooks Range suggests that summer temperature was an important driver of RTS activity over the past 6000 years (Chipman et al., 2016). The temporal trends of the Lake NE14 thermo-erosional
record, summarized here as PC axis 1 values (Fig. 4.5), are broadly similar to the Loon Lake record, with enhanced thermo-erosion 2000-3000 cal BP and during the MCA, in contrast to minimal activity during the cool LIA. However, the Lake NE14 record shows only seven RTS episodes over the past 3000 years, compared to 18 at Loon Lake. The climatic difference between these two sites may explain the higher rate of RTS formation at Loon Lake, as we would expect higher frequency of thermo-erosion in the warmer Noatak compared to the cooler Arctic Foothills site if temperature was the main driver of thermo-erosional activity. In addition, fire activity may have also promoted more frequent thermo-erosion in the Noatak compared to low-fire tundra ecoregions such as the Arctic Foothills.

**Fire-history reconstruction from lake-sediment charcoal**

The charcoal record from Loon Lake (Fig. 4.6) shows a mean fire return interval of 140 (97-196) years based on 22 local fires over the past 3000 years. These results are consistent with the mean fire-return interval of 142 (115-174) estimated from paleorecords in the western portion of the Noatak National Preserve (Higuera et al., 2011b). Summer temperature has been identified as the dominant driver of fire activity in Alaskan tundra regions over the past 60 years (Hu et al., 2015; Young et al., 2016), and the temporal variability in the charcoal record from Loon Lake suggests a link between climate and fire activity at centennial to millennial timescales. The two longest intervals with no fires are from 56 to 576 cal BP (520 years) and from 1326 to 1636 cal BP (310 years), broadly coinciding with the LIA and first millennial cooling (Wiles et al., 2008; Barclay et al., 2009), respectively. Conversely, individual fire-return intervals between 826 and 1326 cal BP, which overlap with the warmer intervening MCA, range from 100 to 150 years and are comparable to the fire cycle in the modern-day boreal forest. However, some of the shortest individual fire-return intervals of the record (30-140 years) occurred ~1600-2100 cal BP, during periods of neoglacial advance in Alaska (Solomina et al., 2015; Kaufman et al., 2016) and overall cool summer temperatures in the northern Brooks Range (Boldt et al., 2015). This discrepancy is not surprising because individual charcoal records reflect the stochastic nature of fire ignitions as well as heterogeneity of climate at fine spatial scales. In addition, vegetation composition influences both landscape flammability and available biomass for burning in tundra ecoregions (Rocha et al., 2012), and changes in local vegetation
composition have been related to site-specific fire frequency inferred from other Noatak charcoal records (Higuera et al., 2011a).

To place the Loon Lake record in the context of the regional fire regime, we compare the timing of local fire events at Loon Lake to the composite CHAR record from sites in the western Noatak (Fig. 4.6), the latter of which are interpreted as reflecting changes in area burned (e.g., Marlon et al., 2009; Kelly et al., 2013). These two records are broadly similar. For example, the highest and lowest values of composite CHAR record correspond to the warm MCA and cool LIA, respectively, and are consistent with fire frequency changes at Loon Lake. The composite CHAR record also exhibits moderate to high values ~1600-2300, when fire frequency at Loon Lake shows frequent burning, suggesting that climate conditions in the western Noatak were conducive to burning. High composite CHAR values ~2600-2900 occur within a broad interval of warm summers in northern Alaska (Boldt et al., 2015), although only two fires are recorded at Loon Lake during this time. Nonetheless, the overall consistency between the charcoal record from Loon Lake and the record of area burned in the region suggest that regional climate was likely the dominant control of fire activity over the past 3000 years.

**Linkages between fire and thermo-erosion**

The temporal trends in the Loon Lake thermo-erosional record and higher frequency of RTS activity compared to Lake NE14 may reflect the influence of fire on thermo-erosion. The most striking result of the SEA is the significant response of MS, Ca:K, Ca:Sr, and PC1 values to high-CHAR fires (Fig. 4.7). The improved SEA relationships for high-CHAR fires, compared to all fires occurring within 0.5 to 1 km of the lake, offer compelling evidence for the link between fire and thermo-erosional regimes, because low-severity fires and fires that occur outside the catchment (i.e., low CHAR peak magnitude) are unlikely to facilitate watershed erosion (e.g., Dunnette et al., 2014; Leys et al., 2016). Observations in ice-rich boreal and tundra areas show evidence for post-fire slump formation within burned areas (e.g., Liljedahl et al., 2007; Lacelle et al., 2010; Thienpoint et al., 2013). For example, post-fire soil warming resulted in widespread thermo-erosional activity and ground subsidence within several years of the large Anaktuvuk River fire on the Alaskan North Slope (Liu et al., 2014; Jones et al., 2015). This rapid response of the landscape to fire was likely driven by changes in surface albedo and radiative forcing, which can persist until vegetation cover is restored and the soil thermal regime reestablishes.
equilibrium with climate, usually within a decade (Rocha et al., 2012). However, recent observations in the Noatak suggest that fires do not trigger RTS initiation on an annual basis (Balser et al., 2014), and the most recent fire inferred from the Loon Lake CHAR record (-49 cal BP) is not associated with positive PC1 values (Fig. 4.6). Moreover, the SEA analyses at Loon Lake show no significant relationship between CHAR-inferred fires and MS, Ca:Sr, Ca:K, or PC1 values in the same decade (Fig. 4.7), suggesting that rapid changes in surface albedo and associated changes to the soil thermal regime following fire were likely not the major controls of thermo-erosion at this site over the past 3000 years.

In contrast to rapid post-fire responses, the 20-30 year time lag in the response of the thermo-erosional regime to high-CHAR fires at Loon Lake (Fig. 4.7) may reflect long-term thermal dynamics as ice-rich permafrost soils establish a new equilibrium with climate following fire. Fires can increase the active-layer depths by up to several meters (e.g., Burn, 1998) and promote the formation of taliks (i.e., areas of unfrozen soil) in deep permafrost soils that can persist long after fire occurrence (Yokishawa et al., 2003). Monitoring of soil temperatures in boreal forest revealed continued active-layer deepening ~20-25 years following fire, with thaw depths increasing from ~45 cm to 3 m (Mackay, 1995; Vierek et al., 2008). The influence of fire on soil temperatures is largely controlled by the thickness and thermal conductivity of the remaining organic soil layer (Yokishawa et al., 2003), and thus permafrost soils become more vulnerable to thaw and collapse with increasing fire-induced loss of overlying organic soil horizons (Brown et al., 2015). Repeat burning in flammable tundra regions such as the Noatak may inhibit post-fire organic soil recovery, leading to long-term increases in active-layer thickness. For example, 17% of the area burned in the Noatak from 1950 to 2011 burned more than once, with an average of 13 years between repeat fire events (Rocha et al., 2012). In addition, high-severity fires can dramatically reduce organic layer thickness and alter soil thermal regimes. On the Alaskan North Slope, the Anaktuvuk River Fire combusted ~30% of the organic soil horizon, which represented ~50 years of organic matter accumulation (Mack et al., 2011), suggesting that the re-accumulation of organic soils following severe tundra fires may take several decades.

Fire-driven changes in vegetation composition may also play a role in modifying the thermal regime of permafrost soils over several decades and promoting thermo-erosion in tundra ecosystems. In the Noatak National Preserve and adjacent areas of the Brooks Range, >50% of
the terrain identified as suitable for RTS formation is found in shrub tundra (Balser, 2015). Shrub-dominated sites have relatively low albedo, deeper snow pack, and elevated near-surface ground temperatures compared to non-shrub areas (McFadden et al., 2001; Sturm et al., 2005; Lantz et al., 2013), and the insulation provided by shrubs and deeper snowpack can result in increased active-layer depth (Mackay and Burn, 2002; Lawrence and Swenson, 2011). Fires may be a key mechanism in promoting shrub proliferation in the Arctic, beyond the direct impacts of climate alone (e.g., Racine et al., 2004; Lantz et al., 2013). While vegetation regrowth within a few years of fire may be similar to pre-fire composition (e.g., Bret-Harte et al., 2013), older burned areas in the tundra are characterized by a higher proportion of shrubs and/or enhanced surface greenness compared to recently burned sites (Racine et al., 2004; Rocha et al., 2012; Jones et al., 2013), suggesting that the post-fire transition to increased shrub abundance may take several decades. In the western Noatak National Preserve, vegetation surveys from burned and unburned areas <10 km from Loon Lake show a significant increase in percent cover of shrubs and sedges from 1982 to 2005 at burned sites (Racine et al., 2006). Such decadal-scale vegetation responses to fire may thus play a role in the lead-lag relationship between fires and thermo-erosional activity at Loon Lake, highlighting the potential for complex interactions between climate, vegetation, and disturbance as tundra ecosystems respond to climate warming.

An alternative interpretation of the lagged response of RTS activity to fire at Loon Lake is that the CHAR and geochemical records reflect rapid versus delayed climate-driven responses of fire versus ground-ice thaw, respectively. Tundra burning in Alaska is primarily a function of temperature and precipitation, and the probability of fire rapidly increases if conditions exceed climatic thresholds to burning (Hu et al. 2015, Young et al. 2016). This relationship was dramatically illustrated by the Anaktuvuk River Fire on the Alaskan North Slope, which occurred in response to anomalously warm and dry conditions in the summer of 2007 (Jones et al., 2009; Hu et al. 2010). In contrast, the response of permafrost to climate change may be delayed because of the lagged propagation of heat from the surface through deeper permafrost soils (Lachenbruch and Marshall, 1986) and the presence of a transient layer between the active-layer and permafrost that inhibits the thaw of deeper ground ice in response to short-term climate variability (Shur et al., 2005). Aerial and satellite imagery from Alaskan tundra areas without fire document rapid and sustained degradation of ice-rich terrain to climate warming over the past few decades (e.g., Jorgenson et al., 2006; Raynolds et al., 2014), suggesting that the response of
permafrost soils to climate can be rapid if conditions are sufficient to dramatically alter the soil thermal regime. However, soils insulated by a thick organic overburden may be less responsive to direct climate-driven changes and likely require disturbance to initiate thermal degradation (Shur and Jorgenson, 2007). It is difficult to gauge the responsiveness of ice-rich soils to changing climate conditions given the limited timescale of the modern climate record and the paucity of long-term monitoring of permafrost thaw depths. While our record from Loon Lake suggests that fire is likely an important factor in altering soil thermal regimes in flammable ice-rich landscapes, additional paleorecords spanning a range of soil characteristics and fire activity are necessary to fully elucidate climate-fire-permafrost relationships.

Additional factors may have influenced the relationship between fire and thermo-erosional activity at Loon Lake over the past 3000 years. The response of ice-rich permafrost terrain to fire is complicated by landscape characteristics that define the thermal regime of the soil, such as soil texture, hydrology, vegetation composition, landscape position, and microclimate (e.g., Swanson, 1996; Nossov et al., 2013; Brown et al., 2015). Such factors may explain why several CHAR peaks identified as fires occurred without evidence for thermo-erosion (e.g., three fires between 2500 and 2650 cal BP; Fig. 4.6). This discrepancy could also be related to spatial heterogeneity in the location of ground-ice deposits and/or variability in the location of burned areas in the watershed, which cannot be assessed with our data. In addition, some RTS episodes occurred during periods with little to no burning. For example, three thermo-erosional episodes are identified from the geochemical records during the LIA (Fig. 4.5), despite overall cool summer temperatures, low regional biomass burning, and a lack of local fires (Fig. 4.6). These episodes could have been driven by variations in moisture variability, which has been linked to modern RTS activity in the Noatak (Balser et al., 2014), and/or reactivation of old slumps in response to positive feedbacks such as talik enlargement under pre-existing RTS features (e.g., Kokelj et al., 2009). Nonetheless, the significant relationship between fire and thermo-erosion at Loon Lake provides new evidence that fires can play an important role in long-term thermo-erosional activity.

Implications for interacting novel disturbance regimes in the Arctic

Disturbance is a key driver of ecosystem processes at a range of spatial and temporal scales (Turner, 2010), and anthropogenic climate change has fundamentally altered natural
disturbance regimes in many ecosystems (e.g., Trapp et al., 2007; Raffa et al., 2008; Dai, 2012, Kelly et al., 2013; Seidl et al. 2014). These novel disturbances can have far-reaching ecological and socio-economic impacts. For example, recent climate-driven increases in wildfires and bark-beetle outbreaks in North American boreal and montane forests suggest that some areas could shift from sinks to sources of atmospheric carbon (Kurz et al., 2008; Chertov et al., 2009; Kelly et al., 2015), thus exacerbating global climate warming. Disturbed terrestrial ecosystems may in turn exhibit heightened sensitivity to ongoing climate change (Kröl-Dulay et al., 2015).

Assessing the role of climate-driven disturbance in ecosystem change is complicated by spatiotemporal variability in disturbance processes as well as interactions between disturbance regimes (e.g., Collins and Smith, 2006; Hicke et al., 2012), which can alter ecosystem resilience and result in unprecedented regime shifts (Paine et al., 1998; Turner, 2010; Buma, 2015; Johnstone et al., 2016). Understanding novel disturbance processes and the interactions between disturbance regimes is thus a critical component of global change studies.

Ongoing climate change in the Arctic may lead to unprecedented fires in tundra areas that have not burned for thousands of years (Chipman et al., 2015; Young et al., 2016), and the impact of these novel fire regimes on permafrost soils is an important component of how Arctic ecosystems respond to climate change. Our records from the flammable Noatak National Preserve provide new insight into the relationship between climate-driven fire activity and thermo-erosional dynamics in ice-rich terrain. These interacting processes can exacerbate the trend of novel disturbances in the Arctic through positive feedbacks and produce cascading effects that amplify the response of Arctic ecosystems to climate change. For example, enhanced thermo-erosion in response to climate and/or fire can increase the sediment load of Arctic lakes and streams beyond climate-driven increases from active-layer thickening and high precipitation (Syvitski, 2002; Mann et al., 2010). Thermo-erosion may also release highly labile DOC stored in ground-ice deposits (Fritz et al., 2015), as well as expose deep carbon reserves to decomposition (Grosse et al., 2011), which can potentially offset the terrestrial carbon gains from climate-driven changes in shrub abundance (Schuur et al., 2009; Kovan et al., 2015) and result in enhanced carbon release to the atmosphere. Increased topographic roughness and changes in surface drainage from thermo-erosion can also facilitate shrub encroachment and increase the amplitude of the carbon cycle (Schuur et al., 2007; Lantz et al., 2009; Belshe et al., 2012). These climate- and disturbance-driven increases in shrub abundance may in turn promote additional
thermo-erosion through changes to the thermal-regime of ice-rich soils (Mackay and Burn 2002; Lawrence and Swenson, 2011), as well as increase flammable biomass, which can result in dramatic changes in tundra fire regimes (e.g., Higuera et al., 2008). Such feedbacks between climate, vegetation, fire, and permafrost disturbance illustrate the vulnerability of tundra landscapes to future climate change, and the importance of understanding these relationships given the rapidly changing state of the Arctic.
Figure 4.2: Sediment magnetic susceptibility (MS), Ca:Sr, and Ca:K from RTS features at Bonus and Loon lakes. Samples from headwall soils are plotted as depth from the top of the exposure. Samples were also obtained from mineral soils exposed at the top and base of the RTSSs, and from organic soils beneath vegetated slump surfaces.
Figure 4.3: Stratigraphic diagram and geochemical proxies from Loon Lake. Elemental data are plotted as normalized counts (raw counts divided by coherent scatter). Bulk density (BD), organic matter (% OM, plotted with reverse x-axis), residual lithic and ash (% lithic) and calcium carbonate (% CaCO₃) were estimated from loss-on-ignition. Magnetic susceptibility (MS) and charcoal concentration (# cm⁻³) are plotted with zoomed x-axis.
Figure 4.4: Age-depth model for Loon Lake. All ages and spline model are plotted with 95% confidence intervals.
Figure 4.5: Lithological and geochemical indicators of thermo-erosion at Loon Lake, with grey bars denoting past RTS episodes. From left to right; sediment bulk density, % CaCO$_3$ from LOI 950 °C, Ca:K, Ca:Sr, magnetic susceptibility (MS), and positive PC axis 1 values (decadal resolution). Columns to the right compare timing of RTS episodes inferred from PC axis 1 values from Loon Lake and Lake NE14 (Chipman et al., 2016).
Figure 4.6: Charcoal-based fire reconstruction at Loon Lake (left), composite charcoal record from Noatak sites with 95% confidence envelope (middle; dimensionless because of standardization), and Loon Lake thermo-erosional records used in SEA (right). Loon Lake charcoal and thermo-erosional records are decadally interpolated and plotted with background trends. Charcoal accumulation rate (CHAR) from Loon Lake shown with peaks interpreted as local fire events (red crosses). Blue dots show fires with low CHAR peak magnitude, interpreted as small or non-catchment fires. Horizontal lines and grey bars show timing of fires and RTS episodes, respectively.
Figure 4.7: Superposed Epoch Analysis (SEA) from Loon Lake showing residual response values (y-axis) of MS, Ca:K, Ca:Sr, and PC axis 1 to all CHAR-inferred fires (left) and high-CHAR fires only (right), plotted with 90, 95, and 99% (MS only) confidence intervals (horizontal blue lines).
REFERENCES


Bowden, W.B., 2010, Climate change in the Arctic - permafrost, thermokarst, and why they matter to the non-Arctic world: Geography Compass, v. 4, p. 1553-1566.

permafrost degradation in Alaskan lowland forests: Journal of Geophysical Research, v. 120, p. 1619-1637.


Swanson, D.K., 2015, Environmental limits of tall shrubs in Alaska’s Arctic national parks, PLOS One, v. 10, p. e0138387.


CHAPTER 5: CONCLUDING REMARKS

Anthropogenic warming may promote rapid permafrost thaw in the Arctic and alter the global carbon cycle. Climate-driven disturbance processes such as fire and thermo-erosion can facilitate rapid ecosystem changes in tundra regions, and may thus exacerbate the direct impacts of ongoing warming. However, the short temporal scale of modern observations, spatial variability of thermo-erosional events, and limited fire occurrence in many tundra ecoregions make it difficult to fully characterize the drivers, variability, and interactions of these processes. This dissertation research utilizes a paleoecological perspective to characterize long-term, natural variability in both tundra fire-regimes and thermo-erosional dynamics, and to assess the interactions between these disturbances over millennia. This research resulted in several key findings, including the long-term persistence of spatial heterogeneity in Alaskan tundra fire-regimes, the association between summer temperature and thermo-erosion over thousands of years, and the potential for decadally-lagged responses of ice-rich terrain to fire.

Paleoecological records provide extended time windows, which are necessary to capture natural variability and provide baselines for assessing modern changes. The new fire reconstructions from historically low-fire Alaskan tundra ecoregions (Chapter 2) are powerful examples of how paleorecords can be used to provide context for modern processes. For example, my research shows that the large Anaktuvuk River Fire on the Alaskan North Slope was unprecedented over the past 6000 years. However, when comparing fire cycles calculated across space and through time, I found that the modern fire regime is not significantly different from the past, and large increases in area burned would be required to define modern burning as unprecedented. This suggests that while individual fires may be anomalous, and can have dramatic impacts on the landscape, detecting changes in the fire regime is only possible with a paleo perspective. This nuance, which is comparable to the difference between weather and climate, is important when considering the impacts of climate change on Arctic ecosystems. Thus, our understanding of modern and future changes in tundra disturbances such as fire will likely benefit greatly from ongoing paleo research in the Arctic.

Permafrost thaw is a powerful driver of geomorphological, ecological, and biogeochemical changes in the Arctic. Ice-rich permafrost areas are extensive in the circumpolar Arctic and are poised to exhibit dramatic changes in response to ongoing climate change. My research (Chapter 3) offered one of the first long-term perspectives on thermo-erosional
dynamics in the tundra. Not only does the record from Lake NE14 reveal that modern thaw slump activity at this site is not unprecedented, it also illustrates the link between summer temperatures and thermo-erosion on millennial timescales. However, this study also highlights the influence of spatial heterogeneity and positive feedbacks in complicating the responses of ice-rich landscapes to climate change. This research is an important contribution to the study of Arctic systems, showcasing the utility of lake sediments for exploring the spatiotemporal dynamics of permafrost disturbance.

Although anecdotal observations of fire-induced thermo-erosion exist from both boreal and tundra ecoregions, there have been no studies documenting how these processes interact over long timescales. For my final dissertation chapter (Chapter 4), I combined paleoecological indicators of fire and thermo-erosion to examine the interactions between tundra disturbance regimes. This research once again illustrates the power of paleoecological methods for assessing ecosystem change, revealing a lagged relationship between fire and thermo-erosional activity that would be difficult to find on the modern landscape given the limited span of observational records and the spatial heterogeneity of ice-rich soils. However, it is important to note that this data is from a single site, and more work is needed to examine these dynamics in other tundra areas. Thus, this study may be a springboard for additional research to determine how different ice-rich tundra landscapes respond to fire regime changes, and to address the mechanistic relationship between these processes.

This dissertation research begs the question: are modern tundra disturbance regimes novel? The answer is that we need more data to fully address this question. For example, while my first study shows no significant difference between past and modern fire regimes, three of the four sites do have a shorter mean fire cycle at present compared to the paleo estimates, which suggest that modern fire regimes may already be changing. Similarly, the thermo-erosional record from Lake NE14 shows repeated thaw slump episodes over the past 6000 years, but the link between summer temperature and thermo-erosion suggests that this process may be enhanced in the future as climate continues to warm. Although the framework for assessing modern disturbance requires additional paleorecords to constrain the past, novel tundra fire regimes are predicted for the future. Thus, understanding the potential interactions between fire and thermo-erosional disturbance is critical for anticipating how these ecosystems will respond to ongoing change. Thermo-erosional responses to both climate and climate-driven fire will
likely vary across space, as suggested by the lack of thermo-erosional activity at Perch Lake (i.e., spatial variability in soils and ground-ice deposits) and heightened thermo-erosional activity at Loon Lake compared to Lake NE14 (i.e., impacts of fire). Elucidating these interactions and their associated variability and feedbacks is a challenging prospect for future research, and will likely benefit from integrative studies that combine paleoecological analyses with both spatial and modelling efforts.
APPENDIX A: SUPPLEMENTARY MATERIAL FOR CHAPTER 2

Table A.1: Radiocarbon ages and $^{210}$Pb activity with modelled ($^{210}$Pb) and calibrated ($^{14}$C) dates from all study sites.

<table>
<thead>
<tr>
<th>Sample depth (cm)</th>
<th>Material</th>
<th>Laboratory ID</th>
<th>210Pb Activity (dpm g$^{-1}$)</th>
<th>$^{14}$C Date (yr BP)</th>
<th>Modeled OR Calibrated Date (cal. yr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Perch Lake</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.00 - 1.00</td>
<td>bulk sediment</td>
<td>UIUC F4-1</td>
<td>30.84 ± 2.8</td>
<td>-58 ± 0.0</td>
<td></td>
</tr>
<tr>
<td>1.00 - 2.00</td>
<td>bulk sediment</td>
<td>UIUC F4-2</td>
<td>23.64 ± 2</td>
<td>-56 ± 1.6</td>
<td></td>
</tr>
<tr>
<td>2.00 - 3.00</td>
<td>bulk sediment</td>
<td>UIUC F4-3</td>
<td>16.33 ± 1.44</td>
<td>-43 ± 2.2</td>
<td></td>
</tr>
<tr>
<td>3.00 - 4.00</td>
<td>bulk sediment</td>
<td>UIUC F4-4</td>
<td>9.24 ± 0.8</td>
<td>-25 ± 2.9</td>
<td></td>
</tr>
<tr>
<td>4.00 - 5.00</td>
<td>bulk sediment</td>
<td>UIUC F4-5</td>
<td>5.88 ± 0.53</td>
<td>-2 ± 3.9</td>
<td></td>
</tr>
<tr>
<td>5.00 - 6.00</td>
<td>bulk sediment</td>
<td>UIUC F4-6</td>
<td>3.15 ± 0.33</td>
<td>28 ± 7.1</td>
<td></td>
</tr>
<tr>
<td>6.00 - 7.00</td>
<td>bulk sediment</td>
<td>UIUC F4-7</td>
<td>2.85 ± 0.27</td>
<td>55 ± 13.0</td>
<td></td>
</tr>
<tr>
<td>7.00 - 8.00</td>
<td>bulk sediment</td>
<td>UIUC F4-8</td>
<td>2.28 ± 0.18</td>
<td>95 ± 27.4</td>
<td></td>
</tr>
<tr>
<td>8.00 - 9.00</td>
<td>bulk sediment</td>
<td>UIUC F4-9</td>
<td>2.04 ± 0.2</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>9.00 - 10.00</td>
<td>bulk sediment</td>
<td>UIUC F4-11</td>
<td>1.88 ± 0.2</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>10.00 - 10.50</td>
<td>bulk sediment</td>
<td>UIUC F4-12</td>
<td>2.05 ± 0.22</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>10.50 - 11.00</td>
<td>bulk sediment</td>
<td>UIUC F4-13</td>
<td>2.1 ± 0.23</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>11.00 - 11.50</td>
<td>bulk sediment</td>
<td>UIUC F4-14</td>
<td>2.08 ± 0.24</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>11.50 - 12.00</td>
<td>bulk sediment</td>
<td>UIUC F4-15</td>
<td>2.07 ± 0.28</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>12.00 - 12.50</td>
<td>bulk sediment</td>
<td>UIUC F4-16</td>
<td>1.81 ± 0.2</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>12.50 - 13.00</td>
<td>bulk sediment</td>
<td>UIUC F4-18</td>
<td>2.06 ± 0.23</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>14.00 - 33.00</td>
<td>twig</td>
<td>CAMS 140576</td>
<td>295 ± 35</td>
<td>382 ± 86</td>
<td></td>
</tr>
<tr>
<td>29.75 - 30.00</td>
<td>twig</td>
<td>CAMS 140577</td>
<td>925 ± 35</td>
<td>847 ± 84</td>
<td></td>
</tr>
<tr>
<td>50.00 - 50.25</td>
<td>wood</td>
<td>CAMS 140275</td>
<td>2460 ± 35</td>
<td>2538 ± 167</td>
<td></td>
</tr>
<tr>
<td>57.50 - 57.75</td>
<td>twig w/bark</td>
<td>CAMS 158760</td>
<td>2735 ± 30</td>
<td>2822 ± 71</td>
<td></td>
</tr>
<tr>
<td>73.25 - 73.50</td>
<td>twigs</td>
<td>CAMS 140578</td>
<td>3560 ± 60</td>
<td>3853 ± 176</td>
<td></td>
</tr>
<tr>
<td>77.00 - 78.00</td>
<td>leaf, wood, roots</td>
<td>CAMS 158761</td>
<td>3805 ± 35</td>
<td>4194 ± 134</td>
<td></td>
</tr>
<tr>
<td>104.00 - 104.50</td>
<td>bark</td>
<td>CAMS 156437</td>
<td>4290 ± 80</td>
<td>4864 ± 302</td>
<td></td>
</tr>
<tr>
<td>129.50 - 130.00</td>
<td>twigs w/bark</td>
<td>CAMS 156439</td>
<td>5915 ± 30</td>
<td>6733 ± 76.5</td>
<td></td>
</tr>
<tr>
<td>167.00 - 167.50</td>
<td>twig w/bark</td>
<td>CAMS 156438</td>
<td>8020 ± 35</td>
<td>8889 ± 131</td>
<td></td>
</tr>
<tr>
<td>216.00 - 216.50</td>
<td>twig w/bark</td>
<td>CAMS 156440</td>
<td>8445 ± 30</td>
<td>9478 ± 45</td>
<td></td>
</tr>
<tr>
<td>Upper Capsule Lake</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>75.00 - 76.00</td>
<td>moss</td>
<td>CAMS 66740</td>
<td>1670 ± 50</td>
<td>1577 ± 135</td>
<td></td>
</tr>
<tr>
<td>100.00 - 101.00</td>
<td>leaves</td>
<td>CAMS 54634</td>
<td>650 ± 80</td>
<td>614 ± 97.5$^d$</td>
<td></td>
</tr>
<tr>
<td>155.00 - 157.00</td>
<td>avg. of 2 dates</td>
<td>---</td>
<td>3705 ± 67</td>
<td>4047 ± 187</td>
<td></td>
</tr>
<tr>
<td>207.00 - 208.00</td>
<td>terrestrial moss</td>
<td>CAMS64013</td>
<td>1070 ± 50</td>
<td>984 ± 109$^d$</td>
<td></td>
</tr>
<tr>
<td>230.00 - 231.00</td>
<td>woody material</td>
<td>CAMS 54635</td>
<td>6610 ± 60</td>
<td>7503 ± 77</td>
<td></td>
</tr>
<tr>
<td>250.00 - 251.00</td>
<td>terrestrial moss</td>
<td>CAMS 66744</td>
<td>8200 ± 50</td>
<td>9161 ± 163</td>
<td></td>
</tr>
<tr>
<td>260.00 - 261.00</td>
<td>leaf, moss,</td>
<td>CAMS 54636</td>
<td>8220 ± 50</td>
<td>9186 ± 174</td>
<td></td>
</tr>
<tr>
<td>310.00 - 311.00</td>
<td>avg. of 6 dates</td>
<td>---</td>
<td>9957 ± 32</td>
<td>11356 ± 163</td>
<td></td>
</tr>
<tr>
<td>325.00 - 326.00</td>
<td>plant and wood</td>
<td>CAMS 54637</td>
<td>9790 ± 50</td>
<td>11214 ± 74.5$^d$</td>
<td></td>
</tr>
<tr>
<td>Keche Lake</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.00 - 0.50</td>
<td>bulk sediment</td>
<td>UIUC P3-1</td>
<td>7.95 ± 0.68</td>
<td>-57 ± 0.0</td>
<td></td>
</tr>
<tr>
<td>0.50 - 1.00</td>
<td>bulk sediment</td>
<td>UIUC P3-2</td>
<td>8.09 ± 0.64</td>
<td>-56 ± 1.9</td>
<td></td>
</tr>
</tbody>
</table>

$^a$ Laboratory ID for samples.
$^b$ Radiocarbon date before the model of Sinclair et al. (2013).
$^c$ Radiocarbon date modelled using Excel VBA software.
$^d$ Radiocarbon date calibrated using OxCal (version 4.3).

---
<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Depth (m)</th>
<th>Method</th>
<th>Material</th>
<th>Depth (m)</th>
<th>Material</th>
<th>Depth (m)</th>
<th>Material</th>
<th>Depth (m)</th>
<th>Material</th>
<th>Depth (m)</th>
<th>Material</th>
<th>Depth (m)</th>
<th>Material</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.00 - 1.50</td>
<td>bulk sediment</td>
<td>UIUC P3-3</td>
<td>6.08 ± 0.4</td>
<td>-55 ± 1.9</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1.50 - 2.00</td>
<td>bulk sediment</td>
<td>UIUC P3-4</td>
<td>4.81 ± 0.26</td>
<td>-53 ± 1.9</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2.00 - 2.50</td>
<td>bulk sediment</td>
<td>UIUC P3-5</td>
<td>4.39 ± 0.33</td>
<td>-52 ± 1.9</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2.50 - 3.00</td>
<td>bulk sediment</td>
<td>UIUC P3-6</td>
<td>4.49 ± 0.39</td>
<td>-51 ± 1.9</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3.00 - 3.50</td>
<td>bulk sediment</td>
<td>UIUC P3-7</td>
<td>4.04 ± 0.32</td>
<td>-50 ± 1.9</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3.50 - 4.00</td>
<td>bulk sediment</td>
<td>UIUC P3-8</td>
<td>4.38 ± 0.32</td>
<td>-48 ± 2.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>4.00 - 4.50</td>
<td>bulk sediment</td>
<td>UIUC P3-9</td>
<td>4.66 ± 0.39</td>
<td>-46 ± 2.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>4.50 - 5.00</td>
<td>bulk sediment</td>
<td>UIUC P3-11</td>
<td>4.40 ± 0.36</td>
<td>-45 ± 2.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5.00 - 5.50</td>
<td>bulk sediment</td>
<td>UIUC P3-12</td>
<td>4.32 ± 0.38</td>
<td>-43 ± 2.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5.50 - 6.00</td>
<td>bulk sediment</td>
<td>UIUC P3-16</td>
<td>5.67 ± 0.51</td>
<td>-41 ± 2.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6.00 - 6.50</td>
<td>bulk sediment</td>
<td>UIUC P3-13</td>
<td>3.65 ± 0.29</td>
<td>-34 ± 2.3</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6.50 - 7.00</td>
<td>bulk sediment</td>
<td>UIUC P3-14</td>
<td>3.66 ± 0.17</td>
<td>-29 ± 2.4</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>7.00 - 7.50</td>
<td>bulk sediment</td>
<td>UIUC P3-15</td>
<td>3.98 ± 0.44</td>
<td>-26 ± 2.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>7.50 - 8.00</td>
<td>bulk sediment</td>
<td>UIUC P3-17</td>
<td>4.02 ± 0.33</td>
<td>-24 ± 2.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>8.00 - 8.50</td>
<td>bulk sediment</td>
<td>UIUC P3-18</td>
<td>3.38 ± 0.3</td>
<td>-21 ± 2.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>8.50 - 9.00</td>
<td>bulk sediment</td>
<td>UIUC P3-19</td>
<td>3.08 ± 0.27</td>
<td>-19 ± 2.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>9.00 - 9.50</td>
<td>bulk sediment</td>
<td>UIUC P3-21</td>
<td>3.22 ± 0.2</td>
<td>-17 ± 2.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>9.50 - 10.00</td>
<td>bulk sediment</td>
<td>UIUC P3-22</td>
<td>5.01 ± 0.44</td>
<td>-15 ± 2.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>10.00 - 10.50</td>
<td>bulk sediment</td>
<td>UIUC P3-23</td>
<td>2.17 ± 0.19</td>
<td>-9 ± 2.4</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>10.50 - 11.00</td>
<td>bulk sediment</td>
<td>UIUC P3-24</td>
<td>2.36 ± 0.17</td>
<td>-7 ± 2.6</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>11.00 - 11.50</td>
<td>bulk sediment</td>
<td>UIUC P3-25</td>
<td>2.49 ± 0.2</td>
<td>-3 ± 2.8</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>11.50 - 12.00</td>
<td>bulk sediment</td>
<td>UIUC P3-26</td>
<td>2.15 ± 0.18</td>
<td>3 ± 3.2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>12.00 - 12.50</td>
<td>bulk sediment</td>
<td>UIUC P3-27</td>
<td>2.38 ± 0.18</td>
<td>5 ± 3.3</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>12.50 - 13.00</td>
<td>bulk sediment</td>
<td>UIUC P3-28</td>
<td>2.30 ± 0.15</td>
<td>11 ± 3.8</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>13.00 - 13.50</td>
<td>bulk sediment</td>
<td>UIUC P3-29</td>
<td>2.33 ± 0.18</td>
<td>17 ± 4.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>13.50 - 14.00</td>
<td>bulk sediment</td>
<td>UIUC P3-30</td>
<td>2.18 ± 0.17</td>
<td>25 ± 5.3</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14.00 - 14.50</td>
<td>bulk sediment</td>
<td>UIUC P3-31</td>
<td>2.27 ± 0.22</td>
<td>31 ± 5.9</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14.50 - 15.00</td>
<td>bulk sediment</td>
<td>UIUC P3-32</td>
<td>1.96 ± 0.2</td>
<td>43 ± 10.2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>15.00 - 15.50</td>
<td>bulk sediment</td>
<td>UIUC P3-33</td>
<td>2.17 ± 0.24</td>
<td>44 ± 10.6</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>15.50 - 16.00</td>
<td>bulk sediment</td>
<td>UIUC P3-34</td>
<td>2.10 ± 0.21</td>
<td>56 ± 16.9</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>16.00 - 16.50</td>
<td>bulk sediment</td>
<td>UIUC P3-35</td>
<td>2.16 ± 0.23</td>
<td>67 ± 20.6</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>16.50 - 17.00</td>
<td>bulk sediment</td>
<td>UIUC P3-36</td>
<td>1.99 ± 0.19</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>17.00 - 17.50</td>
<td>bulk sediment</td>
<td>UIUC P3-37</td>
<td>1.95 ± 0.23</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>17.50 - 18.00</td>
<td>bulk sediment</td>
<td>UIUC P3-38</td>
<td>1.95 ± 0.23</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>18.00 - 18.50</td>
<td>bulk sediment</td>
<td>UIUC P3-39</td>
<td>2.06 ± 0.26</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>18.50 - 19.00</td>
<td>bulk sediment</td>
<td>UIUC P3-40</td>
<td>1.77 ± 0.2</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>19.00 - 19.50</td>
<td>bulk sediment</td>
<td>UIUC P3-41</td>
<td>2.39 ± 0.23</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>19.50 - 20.00</td>
<td>bulk sediment</td>
<td>UIUC P3-42</td>
<td>2.78 ± 0.24</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>20.00 - 20.50</td>
<td>bulk sediment</td>
<td>UIUC P3-43</td>
<td>2.18 ± 0.21</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>20.50 - 21.00</td>
<td>bulk sediment</td>
<td>UIUC P3-44</td>
<td>2.66 ± 0.29</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>21.00 - 21.50</td>
<td>bulk sediment</td>
<td>UIUC P3-45</td>
<td>2.43 ± 0.33</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>21.50 - 22.00</td>
<td>bulk sediment</td>
<td>UIUC P3-46</td>
<td>2.29 ± 0.35</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>22.00 - 22.50</td>
<td>bulk sediment</td>
<td>UIUC P3-47</td>
<td>2.01 ± 0.27</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>22.50 - 23.00</td>
<td>bulk sediment</td>
<td>UIUC P3-48</td>
<td>2.43 ± 0.34</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>23.00 - 23.50</td>
<td>bulk sediment</td>
<td>UIUC P3-49</td>
<td>2.49 ± 0.26</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>23.50 - 24.00</td>
<td>bulk sediment</td>
<td>UIUC P3-50</td>
<td>2.62 ± 0.29</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>24.00 - 24.50</td>
<td>bulk sediment</td>
<td>UIUC P3-51</td>
<td>2.44 ± 0.21</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>24.50 - 25.00</td>
<td>bulk sediment</td>
<td>UIUC P3-52</td>
<td>2.32 ± 0.22</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>25.00 - 25.50</td>
<td>bulk sediment</td>
<td>UIUC P3-53</td>
<td>2.68 ± 0.27</td>
<td>n/a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table A.1 (cont’d)

<table>
<thead>
<tr>
<th>Date</th>
<th>Width</th>
<th>Material Description</th>
<th>Code</th>
<th>14C ± 1σ</th>
<th>210Pb ± 1σ</th>
<th>ID</th>
</tr>
</thead>
<tbody>
<tr>
<td>31.00</td>
<td>32.00</td>
<td>bulk sediment</td>
<td>UIUC P3-54</td>
<td>1.93 ± 0.22</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>41.75</td>
<td>42.00</td>
<td>bark</td>
<td>NSRL-18153</td>
<td>935 ± 20</td>
<td>851 ± 61</td>
<td></td>
</tr>
<tr>
<td>84.25</td>
<td>84.50</td>
<td>2 twigs</td>
<td>NSRL-19851</td>
<td>2485 ± 20</td>
<td>2579 ± 121</td>
<td></td>
</tr>
<tr>
<td>107.00</td>
<td>107.25</td>
<td>wood (no bark)</td>
<td>NSRL-18154</td>
<td>5545 ± 20</td>
<td>6337 ± 49</td>
<td></td>
</tr>
<tr>
<td>112.25</td>
<td>112.50</td>
<td>needle</td>
<td>NSRL-19852</td>
<td>3565 ± 15</td>
<td>3863 ± 36</td>
<td></td>
</tr>
<tr>
<td>128.75</td>
<td>129.00</td>
<td>twig w/bark</td>
<td>NSRL-18155</td>
<td>4015 ± 15</td>
<td>4477 ± 46</td>
<td></td>
</tr>
<tr>
<td>285.25</td>
<td>286.50</td>
<td>twig w/bark</td>
<td>NSRL-19853</td>
<td>8990 ± 20</td>
<td>10194 ± 29</td>
<td></td>
</tr>
</tbody>
</table>

Tungak Lake

<table>
<thead>
<tr>
<th>Date</th>
<th>Width</th>
<th>Material Description</th>
<th>Code</th>
<th>14C ± 1σ</th>
<th>210Pb ± 1σ</th>
<th>ID</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.00</td>
<td>1.00</td>
<td>bulk sediment</td>
<td>UIUC U5-1</td>
<td>18.94 ± 1.34</td>
<td>-62 ± 0.0</td>
<td></td>
</tr>
<tr>
<td>1.00</td>
<td>2.00</td>
<td>bulk sediment</td>
<td>UIUC U5-2</td>
<td>20.7 ± 1.66</td>
<td>-56 ± 0.3</td>
<td></td>
</tr>
<tr>
<td>2.00</td>
<td>3.00</td>
<td>bulk sediment</td>
<td>UIUC U5-3</td>
<td>16.33 ± 1.29</td>
<td>-49 ± 0.4</td>
<td></td>
</tr>
<tr>
<td>3.00</td>
<td>4.00</td>
<td>bulk sediment</td>
<td>UIUC U5-4</td>
<td>8.71 ± 0.76</td>
<td>-21 ± 1.3</td>
<td></td>
</tr>
<tr>
<td>4.00</td>
<td>5.00</td>
<td>bulk sediment</td>
<td>UIUC U5-5</td>
<td>0.748 ± 0.08</td>
<td>30 ± 4.5</td>
<td></td>
</tr>
<tr>
<td>5.00</td>
<td>6.00</td>
<td>bulk sediment</td>
<td>UIUC U5-6</td>
<td>2.057 ± 0.2</td>
<td>30 ± 4.3</td>
<td></td>
</tr>
<tr>
<td>6.00</td>
<td>7.00</td>
<td>bulk sediment</td>
<td>UIUC U5-7</td>
<td>0.707 ± 0.14</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>8.00</td>
<td>9.00</td>
<td>bulk sediment</td>
<td>UIUC U5-8</td>
<td>0.952 ± 0.13</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>10.00</td>
<td>11.00</td>
<td>bulk sediment</td>
<td>UIUC U5-9</td>
<td>0.889 ± 0.08</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>12.00</td>
<td>13.00</td>
<td>bulk sediment</td>
<td>UIUC U5-11</td>
<td>0.847 ± 0.07</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>14.00</td>
<td>15.00</td>
<td>bulk sediment</td>
<td>UIUC U5-12</td>
<td>1.047 ± 0.1</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>16.00</td>
<td>17.00</td>
<td>bulk sediment</td>
<td>UIUC U5-13</td>
<td>1.09 ± 0.11</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>4.50</td>
<td>5.00</td>
<td>tephra</td>
<td>n/a</td>
<td>3430 ± 70</td>
<td>3690 ± 182</td>
<td></td>
</tr>
<tr>
<td>23.00</td>
<td>23.50</td>
<td>leaf and wood</td>
<td>CAMS 160072</td>
<td>9460 ± 150</td>
<td>10753 ± 424</td>
<td></td>
</tr>
<tr>
<td>68.00</td>
<td>68.50</td>
<td>twig w/bark</td>
<td>CAMS 160073</td>
<td>11530 ± 35</td>
<td>13370 ± 101</td>
<td></td>
</tr>
<tr>
<td>208.5</td>
<td>209</td>
<td>wood and bark</td>
<td>CAMS 160074</td>
<td>15240 ± 260</td>
<td>18410 ± 530</td>
<td></td>
</tr>
<tr>
<td>323.50</td>
<td>324.50</td>
<td>bulk sediment</td>
<td>CAMS 161800</td>
<td>27590 ± 50</td>
<td>31622 ± 292</td>
<td></td>
</tr>
</tbody>
</table>

a UIUC: University of Illinois, Urbana-Champaign; CAMS: Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory, Livermore, CA; NSRL: INSTAAR Radiocarbon Laboratory, University of Colorado, Boulder, CO.
b Lead-210 activity, with SD, or conventional radiocarbon years before present (CE 1950), with SD.
c All calibrated dates shown with SD. Bold 210Pb dates were determined using old-age-correction. See Chapter 2 for details on 210Pb modeling and calibration of 14C dates.
d Date omitted from chronology. For details of macrofossils from Upper Capsule Lake, see Oswald et al. (2003).
APPENDIX B: SUPPLEMENTARY MATERIAL FOR CHAPTER 3

STUDY SITES

Figure B.1: Left: Perch Lake (Google™ Earth Image; 30 July 2010). Middle: Lake NE14 with active RTS (black arrow) and blue lake water from sediment influx (aerial photo; 18 July 2007, courtesy of Toolik Field Station http://toolik.alaska.edu/gis/data/index.php). Right: Lake NE14 with clear lake water (Google™ Earth image; 17 July 2009). Coring locations shown with white dots, which correspond to coordinates in Table 3.1.

Figure B.2: Lake NE14 bathymetry and coring site (blue dot). Contour interval = 2 m. Stabilized thaw slump feature shown with hatched fill. Inlets (western and southern shore) and outlet (eastern shore) also shown (arrows denote flow direction). Map created by Toolik Field Station GIS and Remote Sensing Group using bathymetric points obtained with a Garmin GPSMAP 188 Sounder on July 16, 2009. Bathymetric map for Perch Lake is not available.
WATER CHEMISTRY

Table B.1: Modern water chemistry data for Lake NE14. Water chemistry samples were collected between late-June to mid-August of each year. Samples were analyzed at the University of Michigan following methods in Kling et al. (2000).

<table>
<thead>
<tr>
<th>Location/Years</th>
<th>Metric</th>
<th>Cond&lt;sup&gt;a&lt;/sup&gt; μS/cm</th>
<th>Alk&lt;sup&gt;b&lt;/sup&gt; μeq/L</th>
<th>Ca&lt;sup&gt;c&lt;/sup&gt; μM</th>
<th>Mg μM</th>
<th>Na μM</th>
<th>K μM</th>
<th>Si μM</th>
</tr>
</thead>
<tbody>
<tr>
<td>NE14 Lake</td>
<td>mean</td>
<td>146</td>
<td>1476</td>
<td>636</td>
<td>166</td>
<td>17</td>
<td>12</td>
<td>16</td>
</tr>
<tr>
<td></td>
<td>min</td>
<td>142</td>
<td>1444</td>
<td>561</td>
<td>160</td>
<td>14</td>
<td>11</td>
<td>16</td>
</tr>
<tr>
<td></td>
<td>max</td>
<td>159</td>
<td>1536</td>
<td>761</td>
<td>182</td>
<td>23</td>
<td>13</td>
<td>16</td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>NE14 Lake</td>
<td>mean</td>
<td>183</td>
<td>1959</td>
<td>867</td>
<td>216</td>
<td>28</td>
<td>17</td>
<td>21</td>
</tr>
<tr>
<td>2008-2011</td>
<td>std dev</td>
<td>25</td>
<td>182</td>
<td>75</td>
<td>21</td>
<td>3</td>
<td>1</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td>min</td>
<td>127</td>
<td>1677</td>
<td>756</td>
<td>190</td>
<td>25</td>
<td>15</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td>max</td>
<td>321</td>
<td>2544</td>
<td>1105</td>
<td>360</td>
<td>47</td>
<td>21</td>
<td>67</td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>134</td>
<td>90</td>
<td>114</td>
<td>114</td>
<td>114</td>
<td>114</td>
<td>114</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>T-test&lt;sup&gt;d&lt;/sup&gt;</th>
<th>P-value</th>
<th>&lt;0.0001</th>
<th>&lt;0.0001</th>
<th>&lt;0.00016</th>
<th>&lt;0.0001</th>
<th>&lt;0.00017</th>
<th>&lt;0.0001</th>
<th>n/a</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slump Inlet</td>
<td>mean</td>
<td>726</td>
<td>3843</td>
<td>2644</td>
<td>1611</td>
<td>327</td>
<td>69</td>
<td>54</td>
</tr>
<tr>
<td>2008-2011</td>
<td>std dev</td>
<td>196</td>
<td>1206</td>
<td>715</td>
<td>670</td>
<td>163</td>
<td>22</td>
<td>17</td>
</tr>
<tr>
<td></td>
<td>min</td>
<td>325</td>
<td>298</td>
<td>1215</td>
<td>565</td>
<td>30</td>
<td>27</td>
<td>22</td>
</tr>
<tr>
<td></td>
<td>max</td>
<td>1072</td>
<td>5275</td>
<td>3651</td>
<td>2612</td>
<td>551</td>
<td>96</td>
<td>80</td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>23</td>
<td>21</td>
<td>23</td>
<td>23</td>
<td>23</td>
<td>23</td>
<td>23</td>
</tr>
<tr>
<td>Non-slump Inlets&lt;sup&gt;e&lt;/sup&gt;</td>
<td>mean</td>
<td>205</td>
<td>2134</td>
<td>1008</td>
<td>213</td>
<td>23</td>
<td>7</td>
<td>50</td>
</tr>
<tr>
<td>2008-2011</td>
<td>std dev</td>
<td>37</td>
<td>396</td>
<td>176</td>
<td>35</td>
<td>5</td>
<td>2</td>
<td>66</td>
</tr>
<tr>
<td></td>
<td>min</td>
<td>98</td>
<td>999</td>
<td>445</td>
<td>96</td>
<td>9</td>
<td>1</td>
<td>19</td>
</tr>
<tr>
<td></td>
<td>max</td>
<td>278</td>
<td>2938</td>
<td>1340</td>
<td>277</td>
<td>34</td>
<td>13</td>
<td>486</td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>45</td>
<td>45</td>
<td>46</td>
<td>46</td>
<td>46</td>
<td>46</td>
<td>46</td>
</tr>
</tbody>
</table>

<sup>a</sup>Specific conductivity was corrected to 25 °C using a slope of 2% increase per °C increase.

<sup>b</sup>Alkalinity samples (GF/F filtered) were run on a Radiometer autotitrator either at Toolik Field Station or at the University of Michigan.

<sup>c</sup>Cation concentrations were analyzed on GF/F filtered water by a Perkin-Elmer inductively coupled plasma spectrophotometer.

<sup>d</sup>P-values for a two-tailed t-test assuming unequal variances to compare water chemistry values before versus after the slump event, and values for the slump versus watershed inlets.

<sup>e</sup>Inlet stream samples (eastern shore) enter Lake NE14 from Milake and Yurlake, small lakes located within the watershed of Lake NE14 (see Fig. B.1).

CHRONOLOGIES

Terrestrial macrofossils were obtained from the sediments for radiocarbon dating. Macrofossils were cleaned using an acrylic brush and double-distilled water and examined under a dissection microscope at 100X magnification to assure that sediment particles were removed. All terrestrial macrofossils were treated using the acid base-acid procedure of Oswald et al. (2005) and submitted to the Lawrence Livermore National Laboratory for AMS 14C dating.
Radiocarbon ages were calibrated to years before present (BP = before CE 1950) using the Intcal13 dataset in Calib v.7.0.4 (Stuiver and Reimer, 1993; Reimer et al., 2013). The uppermost sediments at Perch Lake were processed following Eakins and Morrison (1978) and $^{210}$Po activity was measured on an Ortec OctetePlus alpha spectrometer at the University of Illinois. Supported $^{210}$Pb activity was calculated following Binford (1990) and a $^{210}$Pb chronology was developed using a constant rate of supply model. The deposition of thermo-erosional material in the uppermost sediments of Lake NE14 precluded $^{210}$Pb analysis at this site.

Final age-depth models based on the $^{210}$Pb and calibrated $^{14}$C dates were created using the classical CLAM age-modeling routine (Blaauw, 2010) in R. We used the smoothing spline function (smoothing = 0.4) with 10,000 bootstrapped iterations. At Perch Lake, all radiocarbon ages were in chronological order and all were used in the final age-depth model (Fig. B.3A). We used sediments from Perch Lake spanning the past 6000 years (Fig. B.3B) for comparison with Lake NE14. At Lake NE14, we omitted two of the 13 calibrated dates because they were obtained from sediments that are interpreted as thermo-erosional events and thus are likely reworked terrestrial material (Fig. B.3C). We also present an adjusted age-depth model for Lake NE14 based on the assumption that the ten identified thermo-erosional episodes throughout the past 6000 years represent instantaneous events (Fig B.3D).

For the NE14 adjusted age model, we assume that the uppermost event (3-7 cm) is dated to -55 cal BP. Annual composites of Landsat imagery indicate that the Lake NE14 RTS released fine sediments to the water column beginning in 2005 (Google Earth Engine 1984-2012 time-lapse for Lake NE14; https://earthengine.google.com/timelapse/#v=68.674527,-149.626373,12,latLng&t=0.20). Lake-water color indicates abundant clay particles in the water column in the summers of 2005-2008. Lake water became clear in summer 2009. We do not know if there was much clay deposition in 2006-2008, because much of the clay band at 3-7 cm might have formed in 2005 or right after, and it took a few years for clay in the water column to settle down (2006-2008). Thus, we assume that the slump sediments from 3-7 cm deposited between 2005 and 2008. The deposition of the top 3 cm of the NE14 core started no later than 2009 (possibly earlier) and lasted until we cored the lake (July 2011), suggesting that the top 3 cm of sediment were deposited in $\geq$ 2.5 years. Based on this, the post-slump sediment accumulation rate at NE14 is 3 cm / 2.5 years = 1.2 cm yr$^{-1}$. However, the top 3-cm of sediment at NE14 are $\sim$70% water, and thus represent only $\sim$0.9 cm of real sediment accumulation. This
deposition rate is not unusually high for core-top sediments. Age-depth models from lakes often show high sedimentation rate near the top of the core because near-surface sediments have very high water content and have not undergone sediment compression. For example, we cored a nearby site in 2008 (Dimple Lake; 40 km north of NE14), and the $^{210}\text{Pb}$ age at 3 cm was 2005 (Hu et al., 2010), indicating that 1 cm of (watery) sediment deposited each year at this site.

**Figure B.3:** Age-depth relationship for A-B) Perch Lake, C) Lake NE14, and D) Lake NE14 assuming instantaneous deposition of thermo-erosional episodes (“adjusted” model). Models are based on $^{14}\text{C}$ dating of terrestrial macrofossils and $^{210}\text{Pb}$ dating of bulk sediment (Perch Lake only). All $^{14}\text{C}$ dates have been calibrated to years BP (years before CE 1950). The spline interpolant (black line) is shown with 95% confidence envelope based on 10,000 bootstraps (grey lines). All dates plotted with 2σ ranges. White circles are dates excluded from the model.
Table B.2: Radiocarbon ages and Lead-210 activity with modeled (\(^{210}\)Pb) and calibrated (\(^{14}\)C) dates from NE14 and Perch Lake.

<table>
<thead>
<tr>
<th>Sample depth (cm)</th>
<th>Adjusted sample depth (cm)</th>
<th>Material(^a)</th>
<th>Laboratory ID(^b)</th>
<th>(^{210})Pb Activity (dpm g(^{-1})) OR</th>
<th>(^{14})C Date (yr BP)(^c)</th>
<th>Modeled OR Calibrated Date (cal. yr BP) with 2(\delta) range(^d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>NE14</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14.00 - 14.50</td>
<td>10.50 - 11.00</td>
<td>twig w/bark</td>
<td>CAMS 160620</td>
<td>130 ± 35</td>
<td>132 ± 8</td>
<td>8 - 278</td>
</tr>
<tr>
<td>30.00 - 31.00</td>
<td>22.75 - 23.75</td>
<td>mixed material</td>
<td>CAMS 158755</td>
<td>720 ± 180</td>
<td>691 ± 320</td>
<td>1049</td>
</tr>
<tr>
<td>40.00 - 40.50</td>
<td>27.00 - 27.50</td>
<td>mixed material</td>
<td>CAMS 157394</td>
<td>910 ± 70</td>
<td>830 ± 692</td>
<td>934</td>
</tr>
<tr>
<td>47.00 - 48.00</td>
<td>34.00 - 35.00</td>
<td>mixed material</td>
<td>CAMS 158756</td>
<td>1075 ± 50</td>
<td>991 ± 914</td>
<td>1172</td>
</tr>
<tr>
<td>65.00 - 66.00</td>
<td>38.25 - 39.25</td>
<td>mixed material</td>
<td>CAMS 158757</td>
<td>1530 ± 35</td>
<td>1425 ± 1530</td>
<td>1523</td>
</tr>
<tr>
<td>87.00 - 87.50</td>
<td>54.25 - 54.75</td>
<td>mixed material</td>
<td>CAMS 157395</td>
<td>1770 ± 60</td>
<td>1689 ± 1552</td>
<td>1857</td>
</tr>
<tr>
<td>95.00 - 95.50</td>
<td>62.00 - 62.50</td>
<td>mixed material</td>
<td>CAMS 160621</td>
<td>1950 ± 70</td>
<td>1900 ± 1717</td>
<td>2099</td>
</tr>
<tr>
<td>116.00 - 116.50</td>
<td>72.25 - 72.75</td>
<td>mixed material</td>
<td>CAMS 157396</td>
<td>3170 ± 100</td>
<td>3490 ± 3399</td>
<td>3569</td>
</tr>
<tr>
<td>136.50 - 137.00</td>
<td>87.00 - 87.50</td>
<td>mixed material</td>
<td>CAMS 160622</td>
<td>3030 ± 70</td>
<td>3223 ± 3005</td>
<td>3380</td>
</tr>
<tr>
<td>149.50 - 150.00</td>
<td>88.50 - 89.00</td>
<td>mixed material</td>
<td>CAMS 156441</td>
<td>3260 ± 35</td>
<td>3387 ± 3082</td>
<td>3633</td>
</tr>
<tr>
<td>159.50 - 160.50</td>
<td>90.00 - 91.00</td>
<td>mixed material</td>
<td>CAMS 158758</td>
<td>3165 ± 40</td>
<td>3392 ± 3255</td>
<td>3466</td>
</tr>
<tr>
<td>168.50 - 169.50</td>
<td>95.75 - 96.75</td>
<td>mixed material</td>
<td>CAMS 158759</td>
<td>3230 ± 40</td>
<td>3451 ± 3379</td>
<td>3515</td>
</tr>
<tr>
<td>181.00 - 181.50</td>
<td>108.25 - 108.75</td>
<td>mixed material</td>
<td>CAMS 156442</td>
<td>4670 ± 30</td>
<td>5399 ± 5316</td>
<td>5568</td>
</tr>
<tr>
<td>Perch</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.00 - 1.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-1</td>
<td>30.84 ± 2.8</td>
<td>-58</td>
<td></td>
</tr>
<tr>
<td>1.00 - 2.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-2</td>
<td>23.64 ± 2</td>
<td>-55</td>
<td>-55 - -56</td>
</tr>
<tr>
<td>2.00 - 3.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-3</td>
<td>16.33 ± 1.44</td>
<td>-42</td>
<td>-39 - -45</td>
</tr>
<tr>
<td>3.00 - 4.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-4</td>
<td>9.24 ± 0.8</td>
<td>-24</td>
<td>-18 - -30</td>
</tr>
<tr>
<td>4.00 - 5.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-5</td>
<td>5.88 ± 0.53</td>
<td>-1</td>
<td>10.9 - -14</td>
</tr>
<tr>
<td>5.00 - 6.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-6</td>
<td>3.15 ± 0.33</td>
<td>29</td>
<td>55.8 - 2.9</td>
</tr>
<tr>
<td>6.00 - 7.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-7</td>
<td>2.85 ± 0.27</td>
<td>58</td>
<td>97.3 - 18.1</td>
</tr>
<tr>
<td>7.00 - 8.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-8</td>
<td>2.28 ± 0.18</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>8.00 - 9.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-9</td>
<td>2.04 ± 0.2</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>9.00 - 10.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-11</td>
<td>1.88 ± 0.2</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>10.00 - 10.50</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-12</td>
<td>2.05 ± 0.22</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>10.50 - 11.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-13</td>
<td>2.1 ± 0.23</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>11.00 - 11.50</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-14</td>
<td>2.08 ± 0.24</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>11.50 - 12.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-15</td>
<td>2.07 ± 0.28</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>12.00 - 12.50</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-16</td>
<td>1.81 ± 0.2</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>12.50 - 13.00</td>
<td>n/a</td>
<td>bulk sediment</td>
<td>UIUC F4-18</td>
<td>2.06 ± 0.23</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>14.00 - 33.00</td>
<td>n/a</td>
<td>twig</td>
<td>CAMS 140576</td>
<td>295 ± 35</td>
<td>384</td>
<td>288 - 464</td>
</tr>
<tr>
<td>29.75 - 30.00</td>
<td>n/a</td>
<td>twig</td>
<td>CAMS 140577</td>
<td>925 ± 35</td>
<td>850</td>
<td>765 - 925</td>
</tr>
<tr>
<td>50.00 - 50.25</td>
<td>n/a</td>
<td>wood</td>
<td>CAMS 140275</td>
<td>2460 ± 35</td>
<td>2557</td>
<td>2365 - 2707</td>
</tr>
<tr>
<td>57.50 - 57.75</td>
<td>n/a</td>
<td>twig w/bark</td>
<td>CAMS 158760</td>
<td>2735 ± 30</td>
<td>2823</td>
<td>2763 - 2917</td>
</tr>
<tr>
<td>73.25 - 73.50</td>
<td>n/a</td>
<td>twigs</td>
<td>CAMS 140578</td>
<td>3560 ± 60</td>
<td>3854</td>
<td>3650 - 4068</td>
</tr>
<tr>
<td>77.00 - 78.00</td>
<td>n/a</td>
<td>mixed material</td>
<td>CAMS 158761</td>
<td>3805 ± 35</td>
<td>4195</td>
<td>4085 - 4382</td>
</tr>
<tr>
<td>104.00 - 104.50</td>
<td>n/a</td>
<td>bark</td>
<td>CAMS 156437</td>
<td>4290 ± 80</td>
<td>4865</td>
<td>4573 - 5259</td>
</tr>
<tr>
<td>129.50 - 130.00</td>
<td>n/a</td>
<td>twigs w/bark</td>
<td>CAMS 156439</td>
<td>5915 ± 30</td>
<td>6733</td>
<td>6664 - 6796</td>
</tr>
<tr>
<td>167.00 - 167.50</td>
<td>n/a</td>
<td>twig w/bark</td>
<td>CAMS 156438</td>
<td>8020 ± 35</td>
<td>8891</td>
<td>8769 - 9013</td>
</tr>
<tr>
<td>216.00 - 216.50</td>
<td>n/a</td>
<td>twig w/bark</td>
<td>CAMS 156440</td>
<td>8445 ± 30</td>
<td>9478</td>
<td>9435 - 9523</td>
</tr>
</tbody>
</table>

\(^a\)Mixed material consists of small terrestrial fragments of wood, bark, leaf, and root material.
\(^b\)UIUC: University of Illinois, Urbana-Champaign; CAMS: Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory, Livermore, CA.
\(^c\)Lead-210 activity or conventional radiocarbon years before present (CE 1950) with standard deviation.
\(^d\)All modeled and calibrated dates shown with 2-sigma ranges.
\(^e\)Dates not included in the age-depth model.
X-RAY FLUORESCENCE

Elemental composition of the continuous 6000-year sediment sequence from both lakes was obtained using X-Ray Fluorescence (XRF). All cores were split lengthwise in the laboratory and the surface of each core was scraped to provide a smooth surface for XRF analysis. Cores were scanned at 0.25-cm (NE14) and 1.0-cm (Perch) resolution with an Itrax XRF Core Scanner at the University of Minnesota, Duluth. Unconsolidated surface samples from each core were packed into sample holders for scanning. High water content of the top 2 cm at Lake NE14 precluded sample analysis. The XRF peak intensities were measured as counts per second (cps) and normalized to coherent scatter to account for variations in surface scattering and dilution of counts by organic matter.

Figure B.4: Stratigraphic diagrams and XRF elemental composition of A) NE14 and B) Perch Lake sediment cores. Normalized counts are raw counts divided by coherent scatter. Calibrated radiocarbon ages (with 2σ ranges) from terrestrial macrofossils plotted on y-axes.
Figure B.5: Lake NE14 photographs of A) uppermost sediments, displayed in 5-dram snap-cap vial lids and B) continuous core sections with stratigraphic logs (simplified in Fig. 3.3). Sediments spanning 3-7 cm are interpreted as the 2005 RTS.
Figure B.6: Perch Lake continuous core section photographs with stratigraphic logs, simplified in Fig. 3.3.
THERMO-EROSIONAL EPISODES

Table B.3: Summary of the thermo-erosional episodes identified from the NE14 sediments. This table includes the depth and age of the sediments from the original age model (as plotted in Figure 3.4) and the adjusted model that assumes instantaneous deposition of the episodes (‘episode timing’; right column of Figure 3.4).

<table>
<thead>
<tr>
<th>Episode #</th>
<th>Top depth (cm)</th>
<th>Base depth (cm)</th>
<th>Top Age (cal BP)</th>
<th>Base Age (cal BP)</th>
<th>Adjusted Depth (cm)</th>
<th>Adjusted Age (cal BP and 2σ range)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>3.0</td>
<td>7.0</td>
<td>-8</td>
<td>66</td>
<td>3.00</td>
<td>53 (-56 - -47)</td>
</tr>
<tr>
<td>2</td>
<td>24.0</td>
<td>28.0</td>
<td>447</td>
<td>548</td>
<td>20.50</td>
<td>532 (431 - 636)</td>
</tr>
<tr>
<td>3</td>
<td>33.0</td>
<td>39.0</td>
<td>673</td>
<td>817</td>
<td>25.75</td>
<td>755 (658 - 841)</td>
</tr>
<tr>
<td>4</td>
<td>49.0</td>
<td>63.0</td>
<td>1052</td>
<td>1352</td>
<td>36.00</td>
<td>1210 (1116 - 1335)</td>
</tr>
<tr>
<td>5</td>
<td>81.0</td>
<td>87.5</td>
<td>1640</td>
<td>1755</td>
<td>54.25</td>
<td>1712 (1627 - 1817)</td>
</tr>
<tr>
<td>6</td>
<td>100.5</td>
<td>102.5</td>
<td>2067</td>
<td>2123</td>
<td>67.50</td>
<td>2160 (2039 - 2300)</td>
</tr>
<tr>
<td>7</td>
<td>107.0</td>
<td>122.0</td>
<td>2252</td>
<td>2673</td>
<td>72.25</td>
<td>2405 (2259 - 2551)</td>
</tr>
<tr>
<td>8</td>
<td>138.0</td>
<td>158.0</td>
<td>2983</td>
<td>3292</td>
<td>88.50</td>
<td>3251 (3104 - 3330)</td>
</tr>
<tr>
<td>9</td>
<td>164.5</td>
<td>168.0</td>
<td>3575</td>
<td>3805</td>
<td>95.00</td>
<td>3557 (3446 - 3726)</td>
</tr>
<tr>
<td>10</td>
<td>184.5</td>
<td>189.5</td>
<td>5552</td>
<td>6135</td>
<td>117.75</td>
<td>5935 (5764 - 6075)</td>
</tr>
</tbody>
</table>

REFERENCES


APPENDIX C: SUPPLEMENTARY MATERIAL FOR CHAPTER 4

CHRONOLOGIES

At 195.5 cm in the Loon Lake sediment core, there is an abrupt transition to sediments rich in terrestrial plant material (Fig C.1). The charcoal concentrations from 195.5 to 219.5 cm are three to four times higher than the rest of the core, lithic content is substantially reduced (mean = 62% compared to 85% for the remainder of the core), and XRF elemental counts of Sr, Si, K, Ti, Rb, and Fe all show a distinct trough in this interval (Fig. 4.3). We assume that the sediments from 195.5-219.5 cm are organic soil material that deposited instantaneously, and thus interpret the $^{14}$C age from within this unit at 218.5 cm (3109 cal BP, Table C.1) as the age of the sedimentary transition at 195.5 cm. From 219.5-279.5 cm, sediments are clay-rich and mottled with sporadic deposits of terrestrial plant material. The wood fragment from 262.25 cm had a calibrated radiocarbon age of 45,451 cal BP (Table C.1), suggesting either reworked terrestrial material or a sedimentary unconformity. We omit this date from the age model and restrict our analyses to the past 3000 years, when chronological control is robust.

Table C.1: Radiocarbon ages and $^{210}$Pb activity with modelled ($^{210}$Pb) and calibrated ($^{14}$C) dates from Loon Lake.

<table>
<thead>
<tr>
<th>Sample depth (cm)</th>
<th>Material</th>
<th>Laboratory ID</th>
<th>$^{210}$Pb Activity (dpm g$^{-1}$)</th>
<th>$^{14}$C Date (yr BP)$^b$</th>
<th>Modeled OR Calibrated Date (cal. yr BP)$^c$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Loon Lake</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.00 - 1.00</td>
<td>bulk sediment</td>
<td>UIUC D6-1</td>
<td>8.17 ± 0.52</td>
<td>-62</td>
<td></td>
</tr>
<tr>
<td>1.00 - 2.00</td>
<td>bulk sediment</td>
<td>UIUC D6-1</td>
<td>7.86 ± 0.36</td>
<td>-51 ± 2.0</td>
<td></td>
</tr>
<tr>
<td>2.00 - 3.00</td>
<td>bulk sediment</td>
<td>UIUC D6-3</td>
<td>4.30 ± 0.23</td>
<td>-30 ± 5.7</td>
<td></td>
</tr>
<tr>
<td>3.00 - 4.00</td>
<td>bulk sediment</td>
<td>UIUC D6-4</td>
<td>2.98 ± 0.16</td>
<td>-10 ± 10.3</td>
<td></td>
</tr>
<tr>
<td>4.00 - 5.00</td>
<td>bulk sediment</td>
<td>UIUC D6-5</td>
<td>3.08 ± 0.27</td>
<td>6 ± 14.9</td>
<td></td>
</tr>
<tr>
<td>5.00 - 6.00</td>
<td>bulk sediment</td>
<td>UIUC D6-6</td>
<td>2.57 ± 0.15</td>
<td>41 ± 14.2</td>
<td></td>
</tr>
<tr>
<td>6.00 - 7.00</td>
<td>bulk sediment</td>
<td>UIUC D6-7</td>
<td>1.91 ± 0.11</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>7.00 - 8.00</td>
<td>bulk sediment</td>
<td>UIUC D6-8</td>
<td>2.09 ± 0.14</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>8.00 - 9.00</td>
<td>bulk sediment</td>
<td>UIUC D6-9</td>
<td>1.97 ± 0.12</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>10.00 - 11.00</td>
<td>bulk sediment</td>
<td>UIUC D6-12</td>
<td>2.18 ± 0.21</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>12.00 - 13.00</td>
<td>bulk sediment</td>
<td>UIUC D6-14</td>
<td>1.93 ± 0.20</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>14.00 - 15.00</td>
<td>bulk sediment</td>
<td>UIUC D6-16</td>
<td>2.09 ± 0.15</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>63.25 - 63.50</td>
<td>twig w/bark</td>
<td>CAMS 161793</td>
<td>1095 ± 30</td>
<td>1001 ± 61</td>
<td></td>
</tr>
<tr>
<td>133.5 - 133.75</td>
<td>twig w/bark</td>
<td>CAMS 161794</td>
<td>2140 ± 35</td>
<td>2128 ± 150</td>
<td></td>
</tr>
<tr>
<td>171.0 - 171.25</td>
<td>twig w/bark</td>
<td>CAMS 161795</td>
<td>2355 ± 35</td>
<td>2376 ± 85</td>
<td></td>
</tr>
</tbody>
</table>
Table C.1 (cont’d)

<table>
<thead>
<tr>
<th>Depth Range</th>
<th>Type</th>
<th>Sample ID</th>
<th>210 Pb Date</th>
<th>210 Pb Age</th>
<th>137 Cs Date</th>
<th>137 Cs Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>218.50 - 218.75</td>
<td>wood</td>
<td>CAMS 161796</td>
<td>2950 ± 35</td>
<td>3109 ± 109</td>
<td>218.50 - 218.75</td>
<td>wood</td>
</tr>
</tbody>
</table>

*UIUC: University of Illinois, Urbana-Champaign; CAMS: Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory, Livermore, CA*

b Lead-210 activity, or conventional radiocarbon years before present (CE 1950), with SD.

c All modeled and calibrated dates shown with 2-sigma ranges. Bold $^{210}$Pb date was determined using old-age-correction.

d Date omitted from chronology

**CORE LOGS AND PHOTOGRAPHS**

*Figure C.1: Loon Lake continuous core sections with stratigraphic logs, simplified in Fig. 4.3.*