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The Response of High Elevation Wetlands to Past Climate Change, and Implications for the Future

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Abstract

As part of the ongoing investigation of mid-latitude fens in Colorado by University of Denver paleoenvironment researchers, research directed at examining the response of mid-latitude subalpine fens to past climatic events potentially analogous to modern anthropogenic climate change was undertaken. As part of this research, a 4.18 m sediment core was retrieved from Harbison Pond (2651 m elev) near Grand Lake, CO and a 3.12 m peat core was collected from Whiskey Fen (2792 m elev) in the Never Summer Range. Stratigraphic evidence indicates the kettle lake formed shortly after the Last Glacial Maximum, at least 16,201 cal yr BP and that peat began accumulating in Whiskey Fen 12,347 cal yr BP. The timing of the formation of this kettle lake provides new details towards the timing of glacial retreat after the Last Glacial Maximum. The recoverable pollen and microscopic charcoal records from Harbison Pond begin 8,804 cal yr BP and reveals that fire intensity and frequency was much greater in the early Holocene than present. Warmer and drier conditions than present during the Holocene Climatic Optimum decreased fire frequency likely due to a decrease in available fuel. Autochthonous production within Harbison Pond increased during this time, while peat stopped being produced in Whiskey Fen. About 2,000 yr BP, modern climate and vegetation conditions became established in the area.The amount of organic carbon storage in subalpine mid-latitude peatlands may be greatly underestimated by current national models, as Whiskey Fen and two other surveyed fens contained thousands of more tons of organic carbon than predicted. Projected anthropogenic climate change will have a larger impact on natural systems connected to mid-latitude subalpine fens than is currently being anticipated.

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The Response of High Elevation Wetlands to Past Climate Change, and Implications for the Future

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ABSTRACT

As part of the ongoing investigation of mid-latitude fens in Colorado by University of Denver paleoenvironment researchers, research directed at examining the response of mid-latitude subalpine fens to past climatic events potentially analogous to modern anthropogenic climate change was undertaken. A 4.18 m sediment core was retrieved from Harbison Pond (2651 m elevation) near Grand Lake, CO and a 3.12 m peat core was collected from Whiskey Fen (2792 m elevation) in the Never Summer Range. Stratigraphic evidence indicates the kettle lake formed shortly after the Last Glacial Maximum, at least 16,201 cal yr BP and that peat began accumulating in Whiskey Fen 12,347 cal yr BP. The timing of the formation of this kettle lake provides new details towards the timing of glacial retreat after the Last Glacial Maximum. The recoverable environmental records from Harbison Pond begin 8,804 cal yr BP and reveals warmer and drier conditions than present during the Holocene Climatic Optimum and decreased fire frequency. Autochthonous production within Harbison Pond increased during this time, while peat stopped being produced in Whiskey Fen. The amount of organic carbon storage in subalpine mid-latitude peatlands may be greatly underestimated by current national models. Projected anthropogenic climate change will have a larger impact on natural systems connected to mid-latitude subalpine fens than is currently being anticipated.

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1. INTRODUCTION

This dissertation represents an attempt to reconstruct changes in the vegetation, climate, and fire regime in the Kawuneeche Valley of central Colorado through the Holocene based on sediment records preserved in a high elevation lake and wetland, and to examine what those changes may mean for conditions in the region in the face of future climate change (Figure 1.1). This research examines changes in the climate of the central Southern Rocky Mountain region since the end of the Pleistocene, approximately 12,000 years ago. Sediment cores were collected from a small lake on a moraine in the Kawuneeche Valley and a wetland on the eastern slope of the Never Summer Range in the Arapahoe National Forest. I employ a multi-proxy approach in analyzing past environmental conditions using these peat and mineral sediment cores to understand past environmental change in the region. These results may also serve to anticipate the effects of future environmental change, and to assist management and preservation efforts. I focus on three important periods in this work, the Late Pleistocene-Early Holocene transition, the Middle Holocene Climatic Optimum, and the Late Holocene. These divisions of the Pleistocene and Holocene follow the International Commission on Stratigraphy's (ICS) global chronostratigraphical correlation table for the last 2.7 million years and the recommendations of the ICS Quaternary subcommittee's Working Group on the Subdivision of the Holocene (ICS, 2016; Walker *et al*., 2012). I use pollen and microscopic charcoal analysis, peat humification, magnetic susceptibility and biogeochemical analyses on cores from the sites to determine how the vegetation and climate of this region have changed since the Pleistocene, how wetland hydrology has responded to climate fluctuation, and how fire frequency has changed as a result of changes in vegetation, climate, and hydrology.

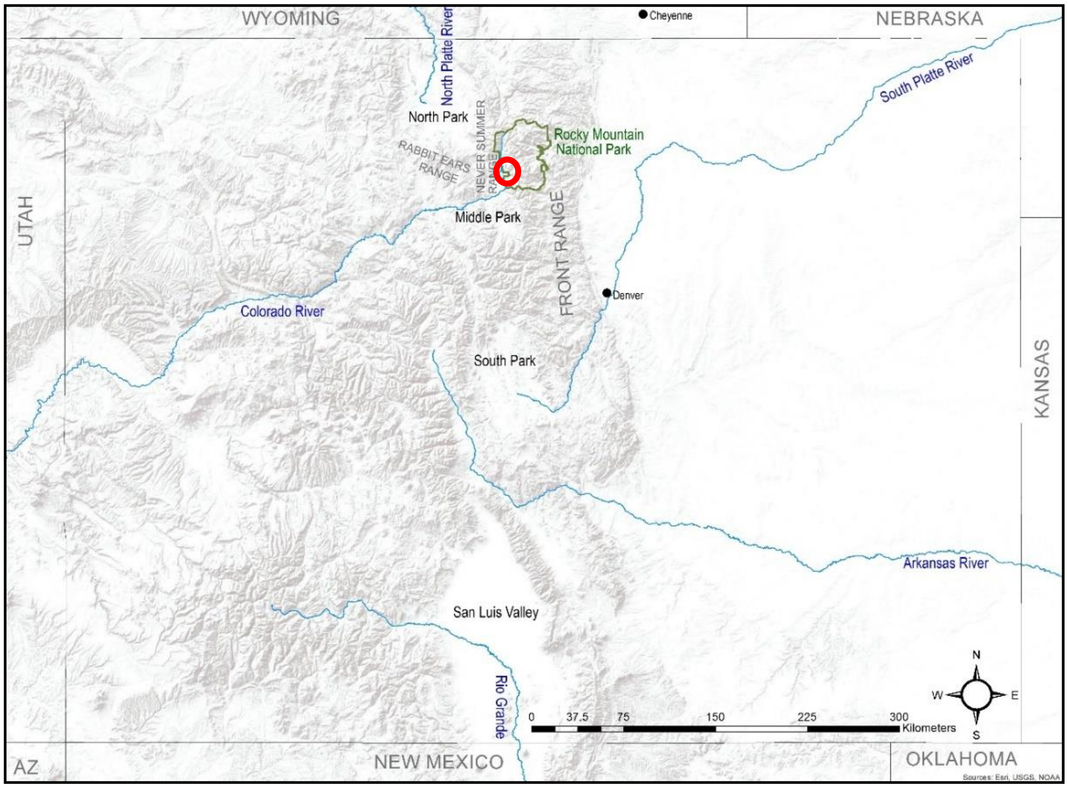

Figure 1.1. Physiographic map of Colorado. The red circle marks the location of Grand Lake and the study sites.

The objectives of my research are to: 1) reconstruct vegetation and climatic changes through the Holocene, 2) reconstruct changes in sedimentation patterns, sedimentation sources, and in situ hydrology and 3) use the information from the first two objectives to assess the scope of possible future impacts from anthropogenic climate change to wetland function, water quality, and water availability in western rivers. I will accomplish these objectives by examining the pollen, magnetic susceptibility, bulk density, organic content, microscopic charcoal, and macroscopic charcoal in the sediment core, and peat humification, bulk density, and organic content in the peat core.

Based on the three objectives above, my research questions are:

1. How stable are modern vegetation communities in the central Southern Rockies?
2. To what extent and in what way is evidence from past climatic events preserved in the record?
3. How have subalpine fens responded to variations in global temperature in the past?
4. How might changes in mid-latitude subalpine fens affect water resources, as well as carbon storage and sequestration?
5. Are there any analogs for future climates in the record that may be used to predict future vegetation change in the area?
6. To what extent are human impacts on vegetation, such as grazing, fire, or agriculture, detectable in the record?

I chose sites that were at or near modern ecotones in an effort to increase the sensitivity of the pollen record. The lake site used in this study is a small kettle lake west of the town of Grand Lake, hereby designated Harbison Pond. Harbison Pond, at 2651 m (8698 ft), is located near lower treeline. Harbison Pond has no surface inlet or outlet, which increases the local signal of pollen and sediment input. The wetland site that I selected and cored is Whiskey Fen, on the east slope of the Never Summer Range and within the Arapahoe National Forest at 2,842 m elevation (9,324 ft). Whiskey Fen is located 4.8 km west of Harbison Pond and provides a complimentary record for comparison between the sediment and peat analyses within the same local area.

The efficacy of using wetlands and small lakes in alpine and subalpine environments for reconstructing paleoenvironmental change has been established in many places. Pollen accumulation rates at sites in the Rocky Mountains of Colorado were compared to the local atmospheric pollen influx (Fall, 1992). Pollen accumulating in small fens and wetlands in forested settings and above treeline were found to be similar to the atmospheric pollen rain present at the sites. Fall (1992) found that lakes received additional pollen influx from overland flow that distorts the pollen record, making wetlands and small lakes without inflowing streams preferable in upland environments.

A fen is a peat-forming wetland whose principle source of moisture is from springs or groundwater, as opposed to a bog, whose principle source of moisture is from rainfall. The southern Rocky Mountains receive inadequate rainfall to support bogs. The peat-producing wetlands of the central southern Rocky Mountains are all fens, receiving moisture from snow melting at high elevations and then percolating beneath the surface until reaching a depression or collection zone in which a peatland forms. While bogs are typically nutrient-poor and highly acidic due to being a closed hydrologic system, fens have the nutrient and pH characteristics of the groundwater that feeds them and have a surface or subsurface outlet. Productivity is low in bogs due to limited nutrient availability and high acidity, leading to the slow accumulation of peat. The productivity of fens is variable, depending on the geologic setting and climate conditions. In peaty wetlands, productivity occurs in the acrotelm, which is the permeable upper layer that contains the actively growing moss, sedges, and roots of vascular plants. The catotelm contains dense accumulated biomaterial that is no longer growing and has low hydraulic conductivity compared to the living material above it. The saturated, anoxic conditions of the catotelm retard decomposition. Variations in the degree of decomposition within a peat profile occur due to fluctuations in saturation level as peat accumulates over time (discussed in more detail in Chapter 2).

Subalpine wetlands provide many critical services that impact both their immediate mountain settings and the watersheds in which they are present. Wetlands improve water quality and moderate water flow within a basin (Carter, 1996). Wetlands also provide habitat for a variety of wildlife and are areas of high biodiversity (Chimner *et al*., 2010). Peatlands act as a carbon sink by storing carbon from plant material. While high latitude peatlands have been the focus of a great deal of research regarding their response to changing environmental conditions, mid-latitude peatlands have received little previous study.

This dissertation also examines the relationship between climate change, vegetation change, and fire within the region, and uses that information to interpret how subalpine fens respond to climate change. Understanding how these subalpine wetlands developed and how they responded to climate fluctuations in the past will provide an analog for the future of these wetlands as they respond to future climate change, with implications for land and water management. The relationship between climate and carbon storage within a subalpine peatland is also examined. Finally, the age of Harbison Pond provides a new interpretation of the extent of the Pinedale and Bull Lake glaciations in Middle Park.

2. LITERATURE REVIEW

2.1 Paleoenvironmental Reconstructions in the Rocky Mountains

The techniques used to reconstruct paleoenvironments in the Rocky Mountains have included palynological, dendrochronological, and geomorphic studies, as well as charcoal analysis. Precipitation and fire history have been of particular interest to many researchers who work in western North America. While dendrological studies have examined the last 2,000 years of paleoclimate to study high resolution records of precipitation, river discharge, and fire throughout the Rocky Mountains, sediment records collected from Idaho, Wyoming and Colorado provide information about changes in fire frequency and climate in the region. This research is focused on paleoenvironmental changes since the last Glacial Maximum, which occurred at about 19,000 yr BP (Owen *et al*., 2003; Pierce, 2003; Benson *et al*., 2005; Clark *et al*., 2009).

For the purposes of clarity and consistency, I will be using chronology divisions as recommended by the International Commission on Stratigraphy (Figure 2.1) (ICS, 2016; Walker *et al*., 2012). The units of time this literature review focuses on extend from the Late Pleistocene until present. The Late Pleistocene occurred from 120,000 to 11,700 yr BP. The Early Holocene lasted from 11,700 yr BP to 8,200 yr BP. A brief and sudden cooling event that lasted several centuries, known as the Younger Dryas, marks the division between these two ages. The Holocene is divided into three ages (Early, Middle, and Late). A major short-term cooling episode that is reflected in isotopic records from Greenland ice cores, known as the 8,200 yr event, marks the boundary between the Early and Middle Holocene (Walker *et al*., 2012). An aridification event experienced in many regions of the world at about 4,200 yr BP marks the boundary between the Middle and Late Holocene (Walker *et al*., 2012).

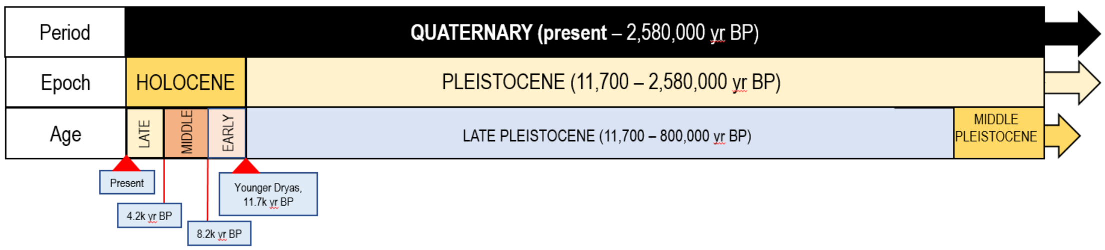

Figure 2.1. Chronology divisions of the Late Pleistocene and Holocene (ICS, 2016; Walker *et al*., 2012).

Paleoenvironmental reconstructions in the Rocky Mountains share a general pattern of a gradual warming trend from cooler than present conditions following deglaciation in the Late Pleistocene. The Younger Dryas, a brief steep decline in temperatures that occurred from about 12,900 to 11,700 yr BP, reversed this trend (Walker *et al*., 2012). The Early Holocene southern Rocky Mountains experienced a climate similar to present, and then transitioned to a warmer and much drier climate in the Early to Middle Holocene (11,700 to 8,400 yr BP). While many records within the southern Rockies contain this Middle Holocene drying, the timing of the onset of these conditions and the length of their occurrence varies (Table 2.1). Warmer conditions during the Middle Holocene are present in paleoclimate records in many parts of the world, and is known as the Holocene Climatic Optimum, or Hypsithermal (Viau *et al*., 2006). Climate in the southern Rockies generally cools from the late-middle Holocene to about 2,000 yr BP, to conditions similar to present. Within this longer cooling trend, a brief period of warming occurred from approximately 1,050 to 850 yr BP that is called the Medieval Climate Anomaly. A subsequent cooling occurred in much of the northern hemisphere called the Little Ice Age from about 500 to 250 yr BP. A description of the trends indicated by studies in or near the southern Rocky Mountains is presented below and summarized in Table 2.1.

Table 2.1. Summary of general climate trends in studies conducted in the southern Rocky Mountain region.

Study and Location	Late Holocene	Middle Holocene	Early Holocene	Late Pleistocene
Benedict *et al.*, 2008; Colorado Rockies	Cool	Warm	-	-
	-	-	-	-
Bright and Davis, 1982; Snake River Plain, Idaho	Cool	Warm	-	-
	Wet	Dry	-	-
Caffrey and Doerner, 2012; Colorado Rockies	-	Warm	-	-
	-	-	-	-
Carrara and McGeehin, 2015; Colorado Rockies	Cool	-	-	-
	-	-	-	-
Doerner and Carrara, 2001; west-central Idaho	Cool	Warm	Warm	Cool
	Wet	Dry	Dry	Dry
Elias *et al.*, 1996; Colorado Rockies	Cool	Cool	Warm	Cool
	-	-	-	-
Fall, 1997; Colorado Rockies	Cool	Warm	Warm	Cool
	-	Wet	Wet	Dry
Lundeen *et al.*, 2013; Southeast Idaho	Cool	Warm	Cool	-
	Wet	Dry	Wet	-
Reasoner and Jodry, 2000; Colorado Rockies	-	-	Warm	Cool
	-	-	-	-
Vierling, 1998; Colorado Rockies	Cool	Warm	Warm	Cool
	Dry	Dry	Wet	Dry
Whitlock *et al.*, 1995	Cool	Warm	Warm	Cool
	Wet	Dry	Dry	Dry

Pollen records from a fen in the Idaho Rocky Mountains indicate deglaciation occurred between 17,000 and 14,000 years BP, with drier than present conditions during the late Pleistocene (Doerner and Carrara, 2001). Climate conditions were cool and moist, similar to present, during the early Holocene in central Idaho. Conditions became significantly warmer and drier than present during the early to middle Holocene, before transitioning to the cooler and more moist conditions that have persisted for the last 2,000 years. Pollen analysis of sediments from sites in the Snake River Plain in Idaho also indicate drier and warmer conditions in the region during the middle Holocene, with a xeric sage steppe transitioning to a mixed steppe from 7,000 yr BP to present (Bright and Davis, 1982).

Records from Wyoming are in close agreement with those from Idaho. By 14,000 yr BP, alpine meadows were present at least 600 m lower in elevation than present in Grand Teton and southern Yellowstone National Parks, and the area had a cooler and somewhat drier climate than present. By the early Holocene, conditions had gradually transitioned to a climate similar to present (Whitlock, 1993). The middle Holocene was warmer and drier than present, with *Picea* and *Abies* taxa giving way to *Pseudotsuga* and *Populus*. This transition may also indicate an increase in fire frequency during the middle Holocene in the Wyoming Rocky Mountains (Whitlock, 1993; Whitlock *et al*., 1995). This middle Holocene drying may have been due to a strengthening subtropical high-pressure system in the eastern Pacific that inhibited summer moisture transport (Whitlock *et al*., 1995). The drier conditions of the middle Holocene in Idaho and Wyoming indicated by pollen records has also been detected in a speleothem stable isotope record from the Wasatch Mountains of southeast Idaho (Lundeen *et al*., 2013). Speleothem records in western North America are more sensitive to changes in winter precipitation, as differences in carbon dioxide concentrations between cave air and drip water are greatest in the winter, which is more favorable to calcite deposition and speleothem growth. This record indicates that a significant reduction in winter precipitation occurred in the region during the middle Holocene (Lundeen *et al*., 2013).

Precipitation shifted towards the summer months as climate warmed in the early and middle Holocene (11,000–4000 yr BP), as indicated by the wide elevational range of *Picea engelmannii* and *Abies lasiocarpa* (Fall, 1997). The shift towards a more summer-dominant precipitation regime is also supported by oxygen isotope records from calcite in northwestern Colorado, which indicate less winter precipitation in the early and mid-Holocene (Anderson, 2011). The reduction in winter precipitation and resulting reduced snowpack likely contributed to the overall drier conditions in the region. Fallen timber recovered from the alpine zone demonstrates that the Colorado Rockies were likely warmer during the mid-Holocene than at present, with upper treeline extending more than 300 m beyond its present elevation (Benedict *et al*., 2008; Carrara and McGeehin, 2015). While subalpine forests in the Colorado Rockies encroached into the alpine zone of the mid-Holocene, the montane forests were becoming mixed with steppe vegetation on the valley floors and in response to drier conditions (Fall, 1997). Relatively moist and cool conditions returned after about 4,000 yr BP until 1,500 yr BP, indicated by a simultaneous increase in nonarboreal taxa and decrease in microscopic charcoal. Climate has been relatively stable over the past 1,500 yr BP, with some evidence of tree establishment expanding above the upper treeline in response to warming over the past 100 years.

An oxygen isotope record from calcite in northwestern Colorado indicates a decline in winter precipitation and a shift towards the majority of precipitation falling in the summer months as climate warmed in the early and middle Holocene (11,000–4,000 yr BP) (Anderson, 2011). The reduction in winter precipitation and resulting reduced snowpack likely contributed to the overall drier conditions in the region. An analysis of pollen records from eight sites on the Colorado Rockies' west slope also indicated a shift towards more summer-dominated annual precipitation as climate warmed in the middle Holocene (Fall, 1997). Upper and lower treeline fluctuations since the late Pleistocene were examined by comparing the pollen record and modern pollen rain at locations, using modern temperature and precipitation lapse rates to infer what those environments were like in the past when those boundaries shifted. Between 15,000 and 11,000 yr BP, upper treeline was 300–700 m lower than present, with a relatively wetter and cooler (2–5°C inferred) climate (Fall, 1997). The subalpine forest expanded upslope after 11,000 yr BP due to warming temperatures, and the increased occurrence of *Picea engelmannii* during the early and mid-Holocene (11,000–4,000 yr BP) may be an indication that there was more summer precipitation than winter precipitation during this time.

Pollen records from these western slope sites showed the middle Holocene to be warmer than today, with upper treeline extending more than 300 m beyond its modern elevation (Fall, 1997). The wide elevational range of *Picea engelmannii* and *Abies lasiocarpa* suggests higher summer precipitation and warmer temperatures on the western slope of the Rockies during the mid-Holocene than today. By 6,000 yr BP, the lower montane forest was mixed with steppe vegetation and the lower boundary of the subalpine forest was thinning, indicating a general drying despite still higher than present precipitation. Between 4,000 and 2,600 yr BP, *Picea engelmannii* expanded downslope from the subalpine forest and the montane forest retreated upslope, in response to drying. Since 4,000 yr BP, the extent of both tundra and steppe vegetation, above upper treeline and below lower treeline respectively, has expanded. The modern climate has been relatively stable over the past 2,000 yr BP with brief episodes of warming, and some evidence of tree establishment expanding above the upper treeline in response to warming over the past 100 years.

Sensitive pollen records from cores collected in the San Juan Mountains and the Front Range suggest that treeline responds rapidly to past temperature changes, including the Younger Dryas climate event (Reasoner and Jodry, 2000). Environmental reconstructions indicate that treeline moved upslope from 13,600 to 12,900 yr BP as climate warmed, before declining due to cooler temperatures during the Younger Dryas from 12,900 to 11,700 yr BP. Treeline again began moving upslope through the Holocene.

A 12,000-year pollen record from Lost Park, a wet meadow east of the northern lobe of South Park, provides further evidence of drying conditions during the mid-Holocene (Vierling, 1998). During the early Holocene, the pollen assemblage was similar to modern observed pollen assemblages near Crested Butte, CO and in northwest Wyoming. These locations are sage (*Artemisia*) steppe ecosystems, and indicate somewhat cooler conditions than present at Lost Park at the start of the Holocene, transitioning to temperatures similar to today from 12,000–9,000 yr BP. *Artemisia* pollen began to decline and *Poaceae* pollen began to increase starting at about 9,100 yr BP, suggesting a shift towards summer-dominant, monsoonal precipitation. By the mid-Holocene (6,000–4,000 yr BP), an increase in microscopic charcoal suggests drier conditions returning. Cooler conditions similar to modern climate began to form beginning at about 1,800 yr BP.

The earliest reconstructed temperatures from insect fossils from sites throughout the Rockies show that late glacial warming began at the latest by 12,800–13,200 yr BP, with summer temperatures only 3–4°C cooler than present but lingering cold winters 19–21°C than present (Elias, 1996; Elias, 1985; Elias, 1983). While insects can disperse from neighboring vegetation zones, changes in the ratio of insect species adapted for one environment relative to insect species adapted to another can be used to infer environmental change over time. Elias examined the ratio of tundra/subalpine forest insect species in sediments along altitudinal transects all within the tundra/subalpine forest ecotone, from sites in Montana, Wyoming, and Colorado. A higher ratio of subalpine forest insect taxa at the sites during the early to mid-Holocene (9,000–3,500 yr BP) indicates higher treeline and warmer conditions than present. Conditions generally cooled from 3,500 yr BP until present. The early Holocene (12,000–9,000 yr BP) was warmer than present and perhaps the warmest period during the epoch. Elias noted that pollen interpretations on the Front Range (Short, 1985) sometimes lagged behind the insect assemblage record, and that the insect assemblage record was consistent with hypothesized mid-Holocene (7,500–5,000 yr BP) dry periods in the region based on archaeological records (discussed further below in Native American History of the Colorado Rockies) (Antevs, 1948; Benedict, 1979). The insect assemblage record shows a gradual cooling after the mid-Holocene, until recent warming.

A sediment core from Bear Lake in Rocky Mountain National Park yielded a 7,000-year record of fire and vegetation change (Caffrey and Doerner, 2012). The lake lies within the subalpine forest on the eastern side of the park, in an area with a record of historic fires. The pollen record indicated a widespread montane forest in the middle Holocene (7,000–3,520 yr BP), with 1.6–5.3 fires per thousand years. A reduced magnetic susceptibility response to fire intensity towards the end of the mid-Holocene suggests that the fires were either not as intense or further from the lake, thus not contributing to increased sediment input from erosion. A decrease in arboreal pollen and charcoal from 3,520–1,760 yr BP indicates a cooling climate with less frequent fire. Peaks in charcoal and an increase in nonarboreal pollen taxa from 1,760–850 yr BP suggest a stand-replacing fire occurred within the basin of the lake during this time, perhaps coincident with the Medieval Climate Anomaly. An increase in arboreal pollen and decrease in charcoal indicates conditions cooled from 850 yr BP to 110 yr BP, before warming and increasing fire activity to present.

A fire and erosional history was constructed from cores collected from Chickaree Lake on the western slope of the Front Range in Rocky Mountain National Park (Dunnette *et al*., 2014). Chickaree Lake is a small, deep lake with ephemeral inlet and outlet streams surrounded by an even-age stand of lodgepole pine (*Pinus contorta*). The lake experienced an average Fire-Return Interval (FRI) of 122 years, with no significant millennial variation in the 6,000 yr record. The principle objective of this study was to examine the long-term biogeochemical effects of fire events on subalpine forests, and the researchers found that the recovery time for carbon and nitrogen availability within the ecosystem following a severe fire was less than the FRI. The authors concluded that if the climate continued to shift towards a warmer and drier state, the resilience of this ecosystem may be degraded or compromised. This finding is in agreement with a separate study that created statistical models of the response of Rocky Mountain fire regimes to climate change, that also suggested that the projected future climate of the region may result in a new fire regime that alters the vegetation communities of the Rockies (Westerling *et al*., 2011).

Fire reconstructions of the past 500 years using dendrochronological methods indicate a strong relationship between drought and fire events (Sibold and Veblen, 2006). Large-scale climate patterns, such as the El Niño-Southern Oscillation and Atlantic Multidecadal Oscillation, have correlative relationship with the occurrence and severity of droughts within the north-central Rockies and likely much of western North America (Donnegan *et al*., 2001; Buechling and Baker, 2004; Sibold and Veblen, 2006). Longer, more coarse fire records from charcoal in lake sediments in the Colorado Rocky Mountains indicate that fire intensity is more responsive to episodic drought conditions than fire frequency is (Higuera *et al*., 2014). Fire frequency over the last 6,000 years was statistically derived from the charcoal record to remain between 150 to 200 years, which is similar to present fire frequency conditions as determined from tree ring analysis. Fire extent and intensity was greater than recent millennia from 6,000 to 2,400 cal yr BP, due to warmer summer temperatures and possibly thicker forests increasing the likelihood of crown fires (Higeura *et al*., 2014; Fall, 1997; Benedict *et al*., 2008).

2.2 Ecology and Disturbance in the Rocky Mountains

The structure of the forests of the Rocky Mountains is not just shaped by shifting climatic envelopes, but also by disturbance regimes (Peet, 1981). Fire, wind, mass movement events, and insects contribute to tree mortality and thus a dynamic cycle of succession. The extent and maintenance of lodgepole pine stands is largely dependent on low-frequency, high-intensity fire events (e.g. Romme and Knight, 1981; Sibold *et al*., 2006). Lodgepole stands are dense and even-aged, with the understories of the trees overlapping at many levels. This structure helps to shade out other tree species, and supports intense stand-replacing fires every 100 or more years. Lodgepole pines have serotinous cones that release seeds when exposed to extreme heat, and the trees themselves grow rapidly compared to competing species in their first few decades of life.

More frequent but less intense fires favor other species, such as ponderosa pine (*Pinus ponderosa*), by reducing ground fuel and the risk of crown fires (Covington and Moore, 1994). Ponderosa pines have evolved thick, resistant bark to withstand ground fires, and to grow in stands with open understories. Fire also alters nutrient availability, which affects forest succession and regeneration (Dunnette *et al*., 2014).

Nitrogen levels within the soil can take 50 to 70 years to return to pre-fire levels in Rocky Mountain subalpine forests. The fire frequency of intense, stand-replacing fires of at least the last 4000 years has allowed sufficient time for nutrient recovery between events, but an increase in the frequency of intense fires in the future may inhibit post-fire recovery and alter the forest stand structures. The frequency of intense wildfires has already been observed to be increasing in western North America since the 1980’s, and projected climate warming in the 21st century will likely continue this trend, creating fire regimes that would likely be incompatible with current forest vegetation (Westerling *et al*., 2006; Westerling *et al*., 2011).

Broad-scale global climate patterns, such as ENSO (El Niño-Southern Oscillation), PDO (Pacific Decadal Oscillation), and AMO (Atlantic Multidecadal Oscillation), have been shown to influence the occurrence of fire in western North America (Donnegan *et al*., 2001; Sibold and Veblen, 2006). Drought conditions are the dominant influence on fire activity and severity within the Rocky Mountains (Buechling and Baker, 2004). Drought conditions are more prevalent in much of western North America during the La Niña phase of ENSO, cool phase of the PDO, and warm phase of the AMO (Sibold and Veblen, 2006). The influence of these climate patterns on fire events is stronger when they occur together. Fire reconstructions of the subalpine forests of the Colorado Rockies have shown that the fire frequency record has been relatively complacent the past 6000 years (Higuera *et al*., 2014). Fire intensity, however, has responded to changes in moisture availability and changes in vegetation density brought about by climate.

Disturbance events also interact with one another (Veblen *et al*., 1994). Blow-downs, in which several trees are knocked down by strong wind, may contribute to increased fire risk (Sibold *et al*.,2007). Such an event also allows for younger trees to establish. Forests that have recently experienced fire are less likely to experience blow-downs, due to this addition of younger trees at the expense of older, more vulnerable ones (Kulakowski and Veblen, 2002). Both fire and blow-downs allow opportunities for new trees to establish, and this increased stand-age heterogeneity can strengthen insect resistance (Mitchell *et* *al*.,1983). Outbreaks of tree-killing insects increase the risk of fire in the short-term, but decrease risk in the long-term (Jenkins *et al*.,2008). The net result of the interactions between fire, wind, and insect events is that the risk of wind and insect events decreases with the frequency of fires with enough intensity to kill trees, and the risk of fire with enough intensity to kill trees increases with wind and insect events. Fire is an unavoidable part of the ecosystem in the forests of the Rockies.

The floor of the southern Kawuneeche Valley (2700 m elevation) and the low country of northern Middle Park around Grand Lake (2551 m elevation) is rich with riparian habitat and wetlands intermixed with low lying ridges and hills forested with lodgepole pine. The valley floor is within the upper montane zone (2450 m–2850 m elevation), but the high water table allows for widespread willow (*Salix* spp.), narrowleaf cottonwood (*Populus angustifolia*), thinleaf alder (*Alnus tenuiflolia*), and sedges (e.g. *Carex aquatilis*), with the occasional Colorado blue spruce (*Picea pungens*) (Peet, 1981). The lower slopes adjacent to the valley floor are dominated by lodgepole pine, with some Douglas fir (*Pseudotsuga menziesii*) and Engelmann spruce (*Picea Engelmanii*) present. Mountain fens occur occasionally on slopes with persistent subsurface water flow from melting snow at higher elevations. The elevation of Middle Park south of Grand Lake remains within the bounds of the lower montane zone (1800–2450 m elevation), but the lower terrain is unforested due to cold air drainage in the winter months inhibiting tree establishment (Peet, 1981). Further upslope, the subalpine zone (2850–3500 m elevation) is almost exclusively composed of Engelmann spruce and subalpine fir (*Abies* *lasiocarpa*). Above approximately 3500 m elevation, tree cover thins and gives way to alpine tundra and permanent snow fields at the highest peaks.

The elevations of these ecotones shift with changing climate conditions through time. Melting ice patches in the Mummy Range have yielded evidence that treeline has fluctuated throughout the Holocene, likely responding to global temperature trends (Benedict *et al*., 2008). The retreating ice patches have revealed the remains of spruce timber at elevations of more than 3465 meters. The wood from these remains were found to have been dead for almost 4000 years, giving them an establishment date of about 4200 years BP. Neoglacial cooling likely led to the death of trees, and increased snow deposition and transport led to the growth of the ice patches that covered preserved the remains of the trees. The ice patch is currently retreating in response to recent warm and dry summers.

2.3 Peat as a Climate Proxy

Peatland establishment has a strong relationship to temperature and aridity. Climate conditions in the early Holocene inhibited peatland establishment in western Canada (Halsey *et al*., 1998). Basal dates from 90 sites in western Canada show that peatland establishment began 8,000–9,000 years BP, as prior to this time seasonal aridity limited moisture availability. Deglaciation in the Alberta foothills and other regions within the Canadian Rockies began 14,000 yrs BP, and pollen evidence reveals expansive grasslands quickly establishing in these areas. In other parts of Canada, the timing of deglaciation, emergence of lowlands due to glacial rebound, and the draining of pro-glacial lakes were the most dominant controls affecting peatland establishment.

An experiment conducted in Ontario attempting to simulate the effects of climate change on fens used water table draw-down as a surrogate for drier conditions due to increased evapotranspiration from warming (Whittington and Price, 2006). A peatland at an experimental site experienced a lowering of the water table by 20 cm. The hydrological response of this site was compared to a control site, as well as another site that had been drained by about 3 m 8 years prior to the experiment. The experimental and drained sites both experienced surface lowering due to compression and increased decomposition. These effects have a positive feedback with one another, amplifying the variation in water table level while decreasing porosity and hydraulic conductivity. Water table variation allows for more decomposition, which then leads to further compression, loss of porosity, and decreased hydraulic conductivity. The outcome of these changes is the decrease in water storage capacity and the transformation from a carbon sink to a carbon source.

A coupled physical-biogeochemical model of peatlands in northern Manitoba found that a peatland’s temperature sensitivity to decomposition increases over time due to interactions between the level of the water table and peat depth (Ise *et al*., 2008). Daily temperatures and water table conditions were modeled at a fen and then compared the results to daily field measurements from the site. Having produced a close approximation of the actual field conditions with the model, Ise *et al*., (2008) added a 4°C temperature increase to a 2,000-year simulation of the site and compared it to a static simulation. The model iteration with a temperature increase showed a 40% loss in soil organic carbon, compared to a 31 % loss of soil organic carbon in the static model. The model was next applied to a nearby peatland with greater peat depth and within a smaller catchment. The model was run under static conditions for 2000 years and produced an output comparable to field parameters. A 4° C temperature increase was applied in year 2000 of the run, and then another 2,000 years were simulated. The rate of soil organic carbon loss was slow in years 2000–2200, with the decomposition of structural organic carbon increasing the input of humic soil organic carbon. After year 2300, the rate of soil organic carbon loss greatly accelerates, culminating in an 86 % loss of soil organic carbon compared to the previous steady-state prior to year 2000. The authors stress that one of the limitations in their model is a lack of ecological and physical responses to simulated environmental changes, which would negatively impact peat accumulation and resistance to decomposition greatly, and could mitigate or exacerbate carbon emissions.

3. REGIONAL SETTING AND STUDY SITE

3.1 Geology and Geomorphology

The Rocky Mountains were uplifted 80 to 55 million years ago when the Laramide orogeny uplifted a broad belt of mountains, which is hypothesized to have been instigated by the dewatering of a subducted slab of oceanic crust (Dickinson *et al*., 1988; Humphreys *et al*., 2003). This process lifted oceanic sedimentary material such as sandstone, siltstone, and shale into high peaks. A second period of uplift occurred between 35 to 25 million years ago, possibly triggered by the subducted slab of oceanic crust fracturing apart (Humphreys, 1995; Karlstrom *et al*., 2012). The Front Range of the Colorado Rockies is composed of high rugged mountains, including 4,346 m Long's Peak. The Never Summer Range is comprised of peaks reaching over 3,600 m elevation. The majority of the uplifted sedimentary material, relatively susceptible to weathering, was eroded and transported to the surrounding plains and lowlands by streams. Erosion exposed older granitic formations on the Front Range that had originated from slow-cooling magma. The ongoing uplift created minor fault zones, sometimes manifesting as linear valleys such as the Kawuneeche. The Rocky Mountains in central Colorado are composed of independent ranges around three intermountain basins: North Park, Middle Park, and South Park (Figure 1.1). North and Middle Park are separated from one another by the Rabbit Ears Range North Park's southern boundary and the Never Summer Range on North Park's eastern boundary. East of the Never Summer Range, the Colorado River flows into and through the Kawuneeche Valley, and drains the Pacific side of the Continental Divide on the west slope of the Front Range, and the east slope of the Never Summer Range. The Colorado River then flows into Middle Park. The Never Summer Range is a geographical oddity, with the eastern slope draining to the Pacific Ocean via the Colorado River and the western slope draining to the Gulf of Mexico by way of the North Platte River through North Park. The Kawuneeche Valley itself is a u-shaped valley that drains the headwaters of the Colorado River, fed by small streams and springs originating from high elevation sources on both sides of the valley (Figure 3.1).

Figure 3.1. Photo of the Kawuneeche Valley from the east side, looking to the south.

3.2 Glacial History

During the last 2.6 million years, the Colorado Rockies have experienced multiple periods of glacial expansion (Pierce, 2003). The Kawuneeche Valley owes it's wide, flat-bottomed form to these past periods of glaciation and subsequent deposition of glacial material. The upper slopes of the mountains along the divide were heavily glaciated during the Pleistocene. Several times during the Pleistocene, large masses of ice flowed from the high peaks of the Front and Never Summer Ranges, eroding the sides and bottom of the Kawuneeche Valley (Richmond, 1974). The precise number of glaciations is unknown, due to the physical evidence of smaller glaciations being erased by more recent, larger ones, but climate reconstructions from deep sea cores indicate approximately 14 major glacial episodes globally over the past 2.6 million years, with more than 20 additional minor episodes (Gibbard *et al*., 2011). Within the context of the study area, the glaciations can be put in three detectable episodic categories: pre-Bull Lake glaciations, the Bull Lake glaciation, and the Pinedale glaciation. These glacial events correspond with marine oxygen-isotope (MIS) stages detected in sea sediments that record global trends temperature through fluctuations in the amount of global continental ice and resulting fluctuations in oxygen-isotope ratios in the shells of some types of marine organisms (Figure 3.2). The stages are numbered backwards from the present, with warm stages being assigned odd numbers and cool stages given even numbers. These Rocky Mountain glaciations roughly correspond to the pre-Illinoian, Illinoian, and Wisconsinan North American continental glaciations.

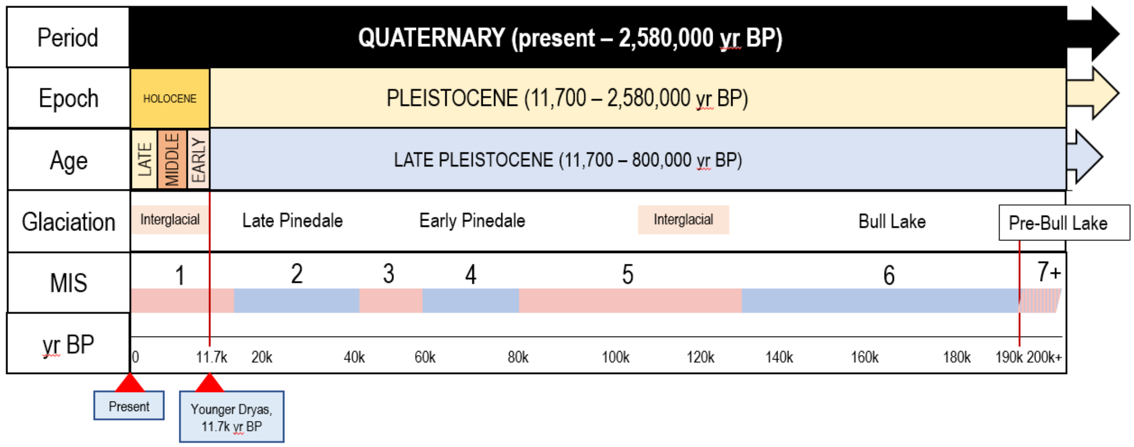

Figure 3.2. Chronology of the Quaternary in the Southern Rocky Mountains. Periods, epochs, ages, and Marine Isotope Stages (MIS) follow the global chronostratigraphic conversion table of the International Commission on Stratigraphy and the recommendations of their Working Group on the Subdivision of the Holocene (ICS, 2016; Walker *et al*., 2012). The timing of glacials and interglacials follows Pierce, 2003.

The Bull Lake (Illinoian) and Pinedale (Wisconsinan) glaciations were named for moraines found on the east and west slopes of the Wind River Range in Wyoming (Blackwelder, 1915). Evidence of at least seven pre-Illinoian glacial events (more than ca. 190,000 years BP) has been identified in stratigraphies analyzed in Montana and Alberta (Karlstrom, 2000), but heavy glaciation of the Colorado Rockies during the Bull Lake and Pinedale glaciations have likely removed most evidence of prior events in the area of the study sites. The Pinedale glaciation is thought to have extended down locally to the moraines that now make up the islands at the south end of Shadow Mountain Reservoir, with the terminal moraine of the Bull Lake glaciation occurring just south of the reservoir (Figures 3.2 and 3.3; Richmond, 1974). Pinedale morainal features are differentiated from Bull Lake features by having thinner, less developed soils, continuity of moraine features, and numerous undrained depressions.

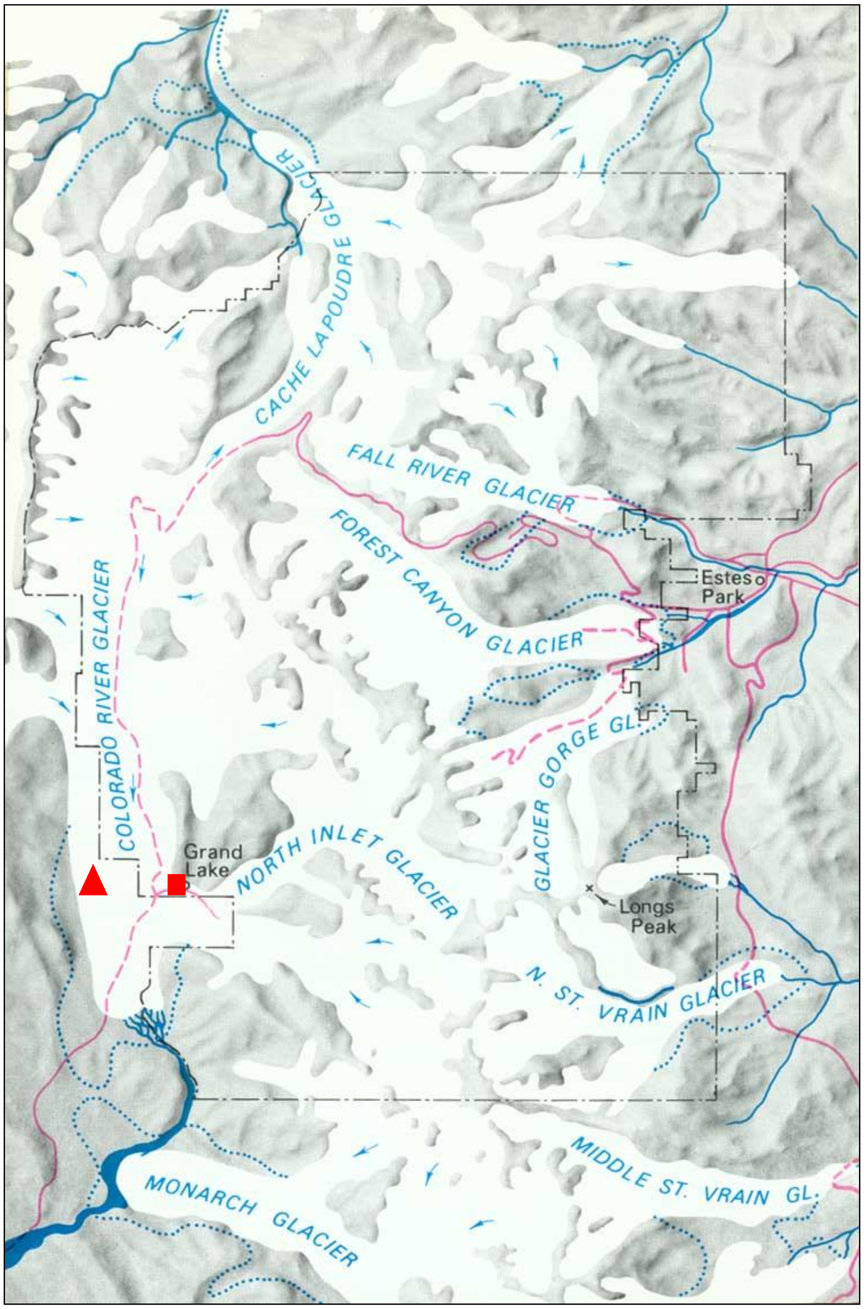

Figure 3.3. Richmond's illustration of glacial extent, from *Raising the Roof of the Rockies, a Geologic History of the Mountains and of the Ice Age in Rocky Mountain National Park* (1974). The blue dotted lines represent the extent of the Bull Lake glaciation, where known. The approximate locations of Harbison Pond (square) and Whiskey Fen (triangle) have been added.

Cosmogenic dating of material at various locations throughout North America from the Pinedale glaciation suggests it reached its peak by about 19,000 years BP, at the Last Glacial Maximum (Owen *et al*., 2003; Pierce, 2003; Benson *et al*., 2005; Clark *et al*., 2009). Dating of glacial features elsewhere in and near the Colorado Rockies indicate that deglaciation after the Pinedale proceeded relatively rapidly after about 17,000 yr BP to 14,000 yr BP, with a small glacial advance occurring during the Younger Dryas about 12,900-11,7800 years ago.

3.3 Climate

The mid-latitude and continental geography of the Colorado Rockies places the area in the path of air masses of multiple origins (Figure 3.4). Mid-latitude cyclones can transport moisture from the north and west from maritime polar air masses, but generally precipitation from this source is light due to the depleted humidity as they move further from the source of the moisture and cross high mountain ranges farther west. Grand Lake, CO receives an average (1981–2010) of 457.5 mm of total precipitation annually (Table 3.1; Arguez *et al*., 2010). The area's average annual temperature is 4.6°C, with an average summer temperature of 13.2°C and an average winter temperature is -1.7°C. Only June, July, and August have average monthly low temperatures above freezing. Grand Lake has a warm summer humid continental (Dfb) Köppen climate type, with areas upslope and within the Kawuneeche Valley having a subarctic (Dfc) Köppen climate type. The highest slopes and peaks in the area have a tundra (ET) Köppen climate type.

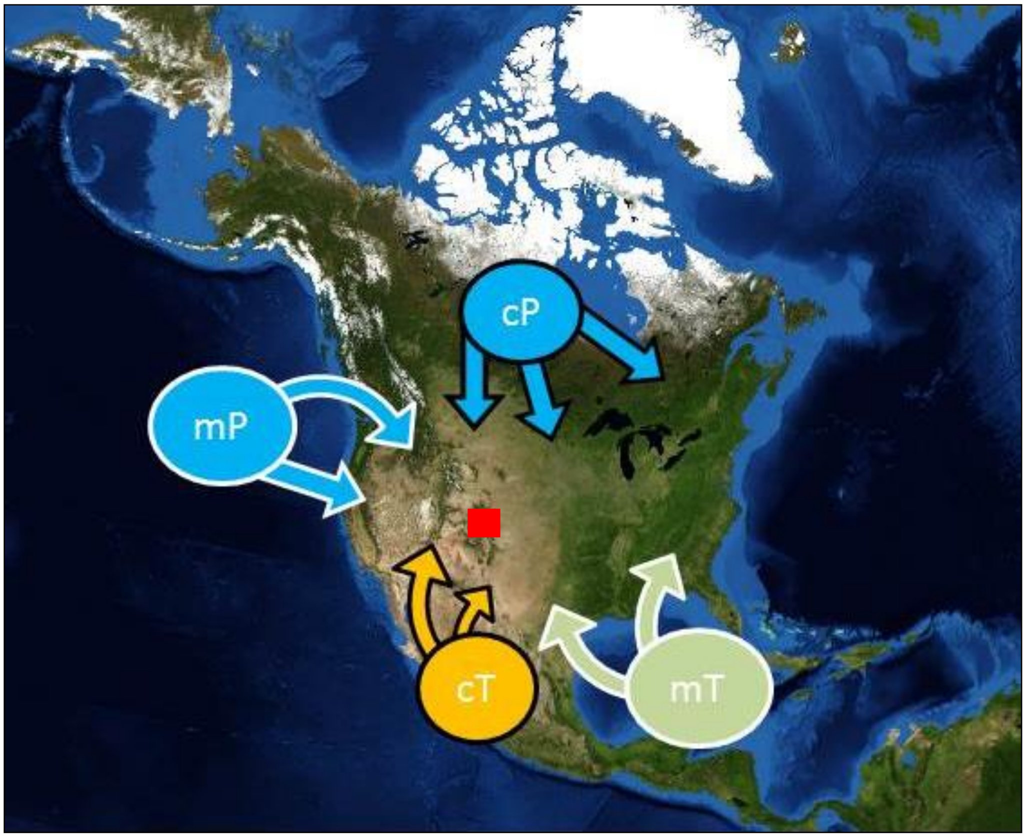

Figure 3.4. Air masses influencing weather patterns of the Colorado Rockies. The red rectangle marks the Colorado Rockies.

Table 3.1. 1981–2010 annual climate data in Middle Park (Arguez *et al*., 2010).

	Hot Sulphur Springs, CO (2341 m elev)	**Grand Lake, CO (2551 m elev)**	**Winter Park, CO (2759 m elev)**
Annual Precipitation	370.6 mm	457.5 mm	490 mm
Mean Winter Precipitation	65.3 mm	68.6 mm	119.1 mm
Mean Summer Precipitation	112.5 mm	115.5 mm	125.2 mm
Average Annual Temperature	4.2°C	4.6°C	2.81°C
Mean Winter Temperature	-4.4°C	-1.7°C	-5.1°C
Mean Summer Temperature	14.1°C	13.2°C	12.1°C

The higher mountain ranges receive enhanced snowfall due to orographic uplift. Polar outbreaks of continental polar air masses from Canada can move south and create larger winter storms after coming into contact with moist air originating from the Gulf of Mexico, but these systems generally affect the plains more than the mountains. The region is fairly dry in the late spring and early summer, with snowmelt providing much of the streamflow and available surface water. The North American monsoon begins to reach northward into the Colorado Rockies in late July, bringing atmospheric instability and moisture that spawn frequent thunderstorms. Drier conditions again return by September and persist until winter.

Oceanic circulation patterns, such as the ENSO and the PDO, have been shown to influence climate patterns in the Colorado Rockies (Gershunov and Barnett, 1998a). ENSO is a sloshing of warm water conditions in the equatorial Pacific that occurs every few years. This pattern has two main phases that each last about a year, along with a neutral phase that typically lasts multiple years. During the neutral phase, sea surface temperatures in the western Pacific are warm while cold water upwelling occurs off the coast of South America in the eastern Pacific. The El Niño phase of ENSO is characterized by a reversal of these sea surface temperatures between the western Pacific and coastal equatorial South America. An El Niño phase is followed by a La Niña phase in which sea surface conditions in the western Pacific are once again warm, but warmer than normal. The PDO is a multidecadal sea surface temperature pattern across the mid-latitudes of the Pacific Ocean. The PDO has two phases, a positive (warm) phase and negative (cool) phase. During the positive phase, the western mid-latitudes of the Pacific Ocean experience warm sea surface temperatures and part of the eastern Pacific experiences cooler than average sea surface temperatures. The negative phase is the reversal of these temperature trends.

ENSO and the PDO affect weather patterns around the world. An El Niño event during the warm phase of the PDO brings increased precipitation to southwestern North America. La Niña events during the cool phase of the PDO are linked to drier conditions in the region (Gershunov and Barnett, 1998b, McCabe and Dettinger, 1999). The variable annual precipitation of western North America is in part a product of these larger global circulation patterns.

3.4 Vegetation

Vegetation patterns and assemblages are formed in large part in response to the climatic envelope of an environment. As that climatic envelope of an area changes over time, so too do the vegetation communities living within that area. The structure of the forests of the Rocky Mountains are also influenced by disturbance regimes (Peet, 1981). Fire, wind, mass movement events, and insects contribute to tree mortality and thus a dynamic cycle of succession. Lodgepole pine (*Pinus contorta*) is one of the dominant tree species in the Colorado Rocky Mountains and is an example of a tree with an ecology that is inextricably linked with fire. The extent and maintenance of lodgepole pine stands is largely dependent on low-frequency, high-intensity fire events (e.g. Romme and Knight, 1981; Sibold *et al*., 2006). Lodgepole stands are dense and even-aged, with the understories of the trees overlapping at many levels. This structure helps to shade out other tree species, and supports intense stand-replacing fires every 100 or more years. Lodgepole pines have serotinous cones that release seeds when exposed to extreme heat, and compared to competing species, lodgepole pines grow rapidly during their first few decades of life.

Elevational effects on temperature and moisture create distinct bands of vegetation communities within mountain environments. Slope-aspect exerts a secondary influence on the location of vegetation zones. The Rocky Mountains in central Colorado is a region of complex topography. Higher elevations experience cooler daily temperatures and thus lower potential evapotranspiration. They also tend to receive more precipitation due to orographic effects. South-facing slopes receive more incoming solar radiation due to the subsolar point of the Earth always falling somewhere to the south, leading to less persistent snow and generally drier conditions. West-facing slopes also experience somewhat higher potential evaporation due to receiving direct sunlight when the atmosphere has already warmed up, but this effect is partly mitigated by being on the windward side of the prevailing winds, as is the case in western United States.

The vegetation types of Middle Park and the Kawuneeche Valley can be broadly divided into seven community types: 1) grasslands and sedge meadows, 2) steppe shrublands, 3) montane forest, 4) lodgepole pine forest, 5) spruce (*Picea*) and fir (*Abies*) forest, and 6) alpine meadow (Peet, 1981). Grasslands and sedge meadows are found in parks and valleys, where grasses and grass-like plants are able to competitively exclude woody plants due to continuous sod, semi-arid conditions, and fine-textured soils further reducing available moisture for tree establishment. Steppe shrublands are found at elevations less than 2,000 m (6,600 ft) on south-facing slopes and rocky soils, and are an open community composed of alder-leave mountain-mahogany (*Cercocarpus montanus*), juniper (*Juniperus scopulorum*), sagebrush (*Artemisia tridentata*) and, increasingly, invasive cheat grass (*Bromus tectorum*). These two community types can transition into montane forest or lodgepole pine forest as elevation increases or slope-aspect allows for higher moisture availability. The most diverse stands of montane forest occur above 2,200 m (7,200 ft) in ravines or floodplains, and are dominated by aspen (*Populus tremuloides*), red birch (*Betula occidentalis*), narrowleaf cottonwood (*Populus angustifolia*), thinleaf alder, and Colorado blue spruce (*Picea pungens*). Uniform stands of lodgepole pine often occur in more xeric areas at the same elevations as montane forest, between 2,000–2,750 m (6,600–9,000 ft). The subalpine forests can be found as low as 2,500 m (8,200 ft) in elevation, but typically occur between 2,750–3,400 m (9,000–11,000 ft). Engelmann spruce (*Picea engelmannii*) and subalpine fir (*Abies* *lasiocarpa*) are two of the most important tree species of the Colorado subalpine forest. Lodgepole pine can also occur within the subalpine forest. In more xeric or exposed sites, limber pine (*Pinus flexilis*) and Rocky Mountain bristlecone pine (*Pinus aristata*) occur. Alpine meadows and tundra become the dominant vegetation on the higher slopes of the Colorado Rockies above 3,400 m (11,000 ft), transitioning away from the subalpine forest completely by about 3,500 m (11,500 ft) (Peet, 1981). This transition is marked by increasingly sparse stands of twisted and stunted subalpine trees between the timberline and treeline. The alpine is composed of various grasses, sedges, and forbs, as well as wildflowers and lichens.

The elevations of these ecotones shift with changing climate conditions through time. Melting ice patches in the Mummy Range have yielded evidence that treeline has fluctuated throughout the Holocene, likely responding to global temperature trends (Benedict *et al*., 2008). The retreating ice patches have revealed the remains of spruce timber at elevations of more than 3,465 meters. Wood from these remains were found to have been dead for almost 4,000 years, giving them an establishment date of about 4,200 years BP. Neoglacial cooling likely led to the death of trees, and increased snow deposition and transport led to the growth of the ice patches that covered preserved the remains of the trees. The ice patch is currently retreating in response to recent warm and dry summers.

Historical photos also provide evidence of recent ecological changes occurring within the Colorado Rockies. Historical photos from the San Juan Mountains show an increase in extent of both conifers and aspen into previous alpine and park areas (Zier and Baker, 2006). Much of this is likely due to forest recovery from disturbance events linked to European settlement. However, conifers were also expanding into wetlands and willow (*Salix*) hummocks, which is not explained by forest recovery from disturbance. Water availability and persistence is an important factor for willow seedling establishment in the Colorado Rockies (Woods and Cooper, 2005).

3.5 Study Site

Harbison Pond is located in north-central Colorado, in the city of Grand Lake at 2,651 m elevation (Figure 3.5; Figure 3.6). The forest around the ponds has been set aside by the city of Grand Lake as a recreational hiking area and contains a network of trails and a main path that leads to the Colorado River. The city was partly motivated to conserve the area due to the presence of a disjunct population of wood frog (*Rana sylvatica*) that is found in the mountains and valleys of north-central Colorado.

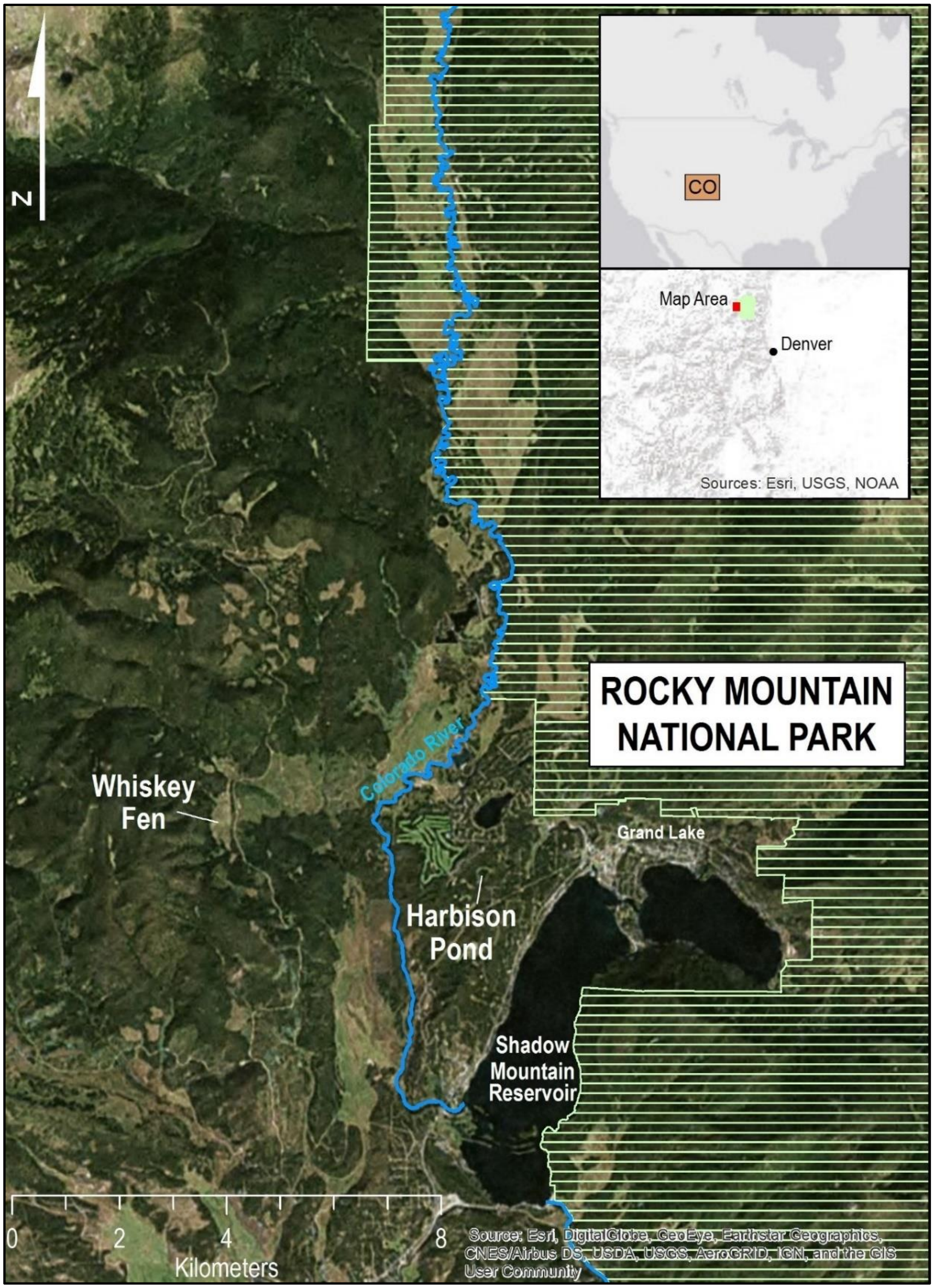

Figure 3.5. Study site area. Harbison Pond lies on recreational land owned by the city of Grand Lake. Whiskey Fen lies within Arapahoe National Forest. The extent of Rocky Mountain National Park is shown in light green.

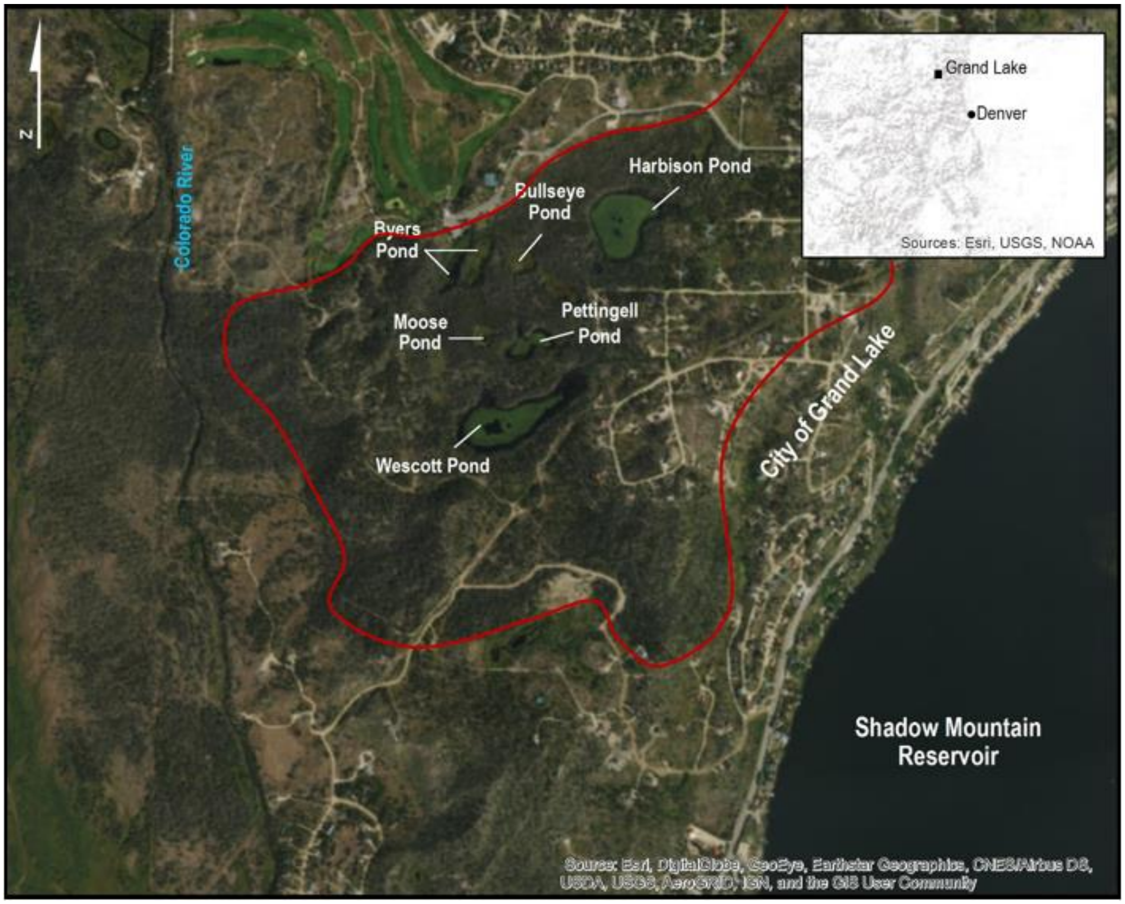

Figure 3.6. Location of kettle ponds on the moraine adjacent to the city of Grand Lake. The morainal material is outlined in red.

Harbison Pond is located 1.2 km northwest of Shadow Mountain Reservoir. Harbison Pond is the largest of a group of small kettle lakes and fens on a large moraine running roughly east-west across the Kawuneeche Valley floor. Kettle lakes are formed from depressions left behind by fragments of retreating glaciers. While some of these kettle lakes have complex shapes that make defining where one begins and another ends somewhat arbitrary, these ponds can be broadly grouped into six water bodies at the time of highest water level annually, with potentially a seventh pond at the southern foot of the moraine that may have been influenced by the construction of an adjacent landfill. These ponds have no surface drainage. The area around the ponds has a thin soil layer (less than 6 cm) with white powdery subsoil and the occasional larger granite boulder. The vegetation on the moraine is a “dog-hair” stand of lodgepole pine.

The surface of the two largest ponds, Harbison Pond and Wescott Pond, are covered with water lilies (*Nuphar polysepala*) except for a 2 to 3-meter-wide apron of open water around their perimeters (Figure 3.6). This apron is also present in Byers Pond, which is of a similar depth as the two larger ponds. This apron may reflect the seasonal or interannual variation in water depth, the zone that freezes to the bottom in the winter, areas too shallow for the aquatic vegetation present, or perhaps a combination of these and other factors. Historical satellite imagery shows small patches of open water existing in the center of Harbison and Wescott Ponds in some years.

The three smaller and shallower ponds are surrounded by sedges that extend into the water, before giving way to open water at no more than 0.5 m of water depth. Two of these smaller ponds were observed to be nearly empty in August of 2014, with the northern end of Bullseye and the southwestern lobe of Pettingell Ponds being the only portions with surface water.

The Colorado River has worked a course through this moraine 1 km to the west of Harbison Pond. At this point the Colorado River exits the Kawuneeche Valley and drains into Shadow Mountain Reservoir before resuming its course through Middle Park. The Kawuneeche Valley is moist, with much of the valley floor occupied by riparian wetlands, abandoned oxbows, and meadows. South of Shadow Mountain Reservoir, conditions become relatively drier with fewer wetlands and ponds. The lodgepole pine stands on the hills and ridges give way to shrubland and grasslands west and south of Shadow Mountain Reservoir, with the exception of riparian corridors.

Whiskey Fen is located 4.83 km west of Harbison Pond on the eastern slope of the Never Summer Range within the Arapahoe National Forest, at an elevation of 2,842 m (9,324 ft). Whiskey Fen extends north to south 1,648 m and is 373 m wide. Its northern edge drains into South Supply Creek, a tributary of the Colorado River. The fen lies adjacent to an access road that sees regular use during hunting season, and the area is popular with local cross-country skiers. The fen was likely formed by a lateral moraine blocking surface drainage.

Both Harbison Pond and Whiskey Fen are located near present-day ecotones. While the Kawuneeche Valley is a moist environment with the valley floor mostly occupied by wetlands and surface water, much of Middle Park is relatively dry and without forest cover. This transition from moist to dry conditions to the south, towards the medial axis of Middle Park, is particularly pronounced at the southern edge of the moraine on which Harbison Pond lies. On slopes, pine stands become increasingly isolated before giving way to sparse juniper and sagebrush. In low lying areas, trees completely give way to grasslands and sedge meadows, while sagebush becomes more abundant. Whiskey Fen lies below the present-day transition between the montane and subalpine forests.

The northern arm of Middle Park, west of Grand Lake and where Harbison Pond is located, was glaciated during the late Pleistocene. Glacial processes during that time have largely created the physical shape of the valley floor. The immediate vicinity around Whiskey Fen shows no evidence of glaciation and was very likely upslope of the

Colorado River Glacier and to the east and south of a tributary glacier coming from the Never Summer Range to the north.

4. FIELD AND LABORATORY METHODS

4.1 Site Selection and Assessment

The initial search and assessment of potential sites in the Colorado River headwaters area focused on the western half of Rocky Mountain National Park. James Doerner, a geographer at the University of Northern Colorado who has done extensive paleoenvironmental work throughout the Rockies and in eastern Rocky Mountain National Park, suggested examining lakes west of the continental divide to fill a need for a reconstructed fire history of the area. Maps and satellite imagery were analyzed for lakes or ponds that appeared to be natural and persistently filled with water in the area. The geology and geomorphology of the region was studied to constrain potential sites to ones likely to have persisted through the Holocene. A map and brief description of visited potential sites can be found in Appendix D.

The cluster of small kettle lakes 0.5 km south of Columbine Lake are surrounded by a lodgepole pine stand. Kettle lakes are formed by large chunks of ice left behind by retreating glaciers creating depressions in loose, glacial till. None of the kettle lakes have any surface inlets or outlets. The water levels in Byers and Moose Ponds fluctuate throughout the year and may sometimes dry up completely. The edges of Pettingell and Wescott Ponds were consistently steep or thickly wooded, making coring access difficult.

Also, local residents recalled that prior to the area becoming protected, all-terrain vehicle motorists would ride through the shallow ends of Pettingell and Wescott Ponds when the water was low. The remaining two ponds, Bullseye and Harbison Pond, were selected for coring. Permission to do so was granted by the city of Grand Lake.

Whiskey Fen is 5 km west of Harbison Pond, adjacent to Colorado Highway 455 in Arapahoe National Forest. South Supply Creek flows through the north end of the fen. Whiskey Fen is the largest wet meadow on the east slope of the Never Summer Range, and has a small camping area at its northern end near South Supply Creek. Peat depth was determined using a 2-m steel probe at various locations throughout the fens. Most of the fen contained at least 1.5 m of peat, with the center of the south end of the fen having more than 2 m of peat.

4.2 Lake Coring

The deepest area of Harbison Pond is 4.3 m (14.1 ft). The center of the lake is free of surface vegetation (Figure 4.1). A coring platform was constructed onshore and towed to the coring site where it was anchored. A 4.18 m core and an overlapping 2.5 m core were recovered using a modified Livingston square-rod corer. For each core site, coring was stopped when unyielding sediment was encountered. The sediment-water interface and uppermost sediment was cored with a clear plastic tube, which was then dispensed in 2 cm increments into plastic sealable bags until stiffer sediment was encounter at 28 cm depth. Each subsequent meter of sediment was extruded, wrapped, and labeled, and then placed in protective tubes and brought to the University of Denver Pollen Analysis Laboratory.

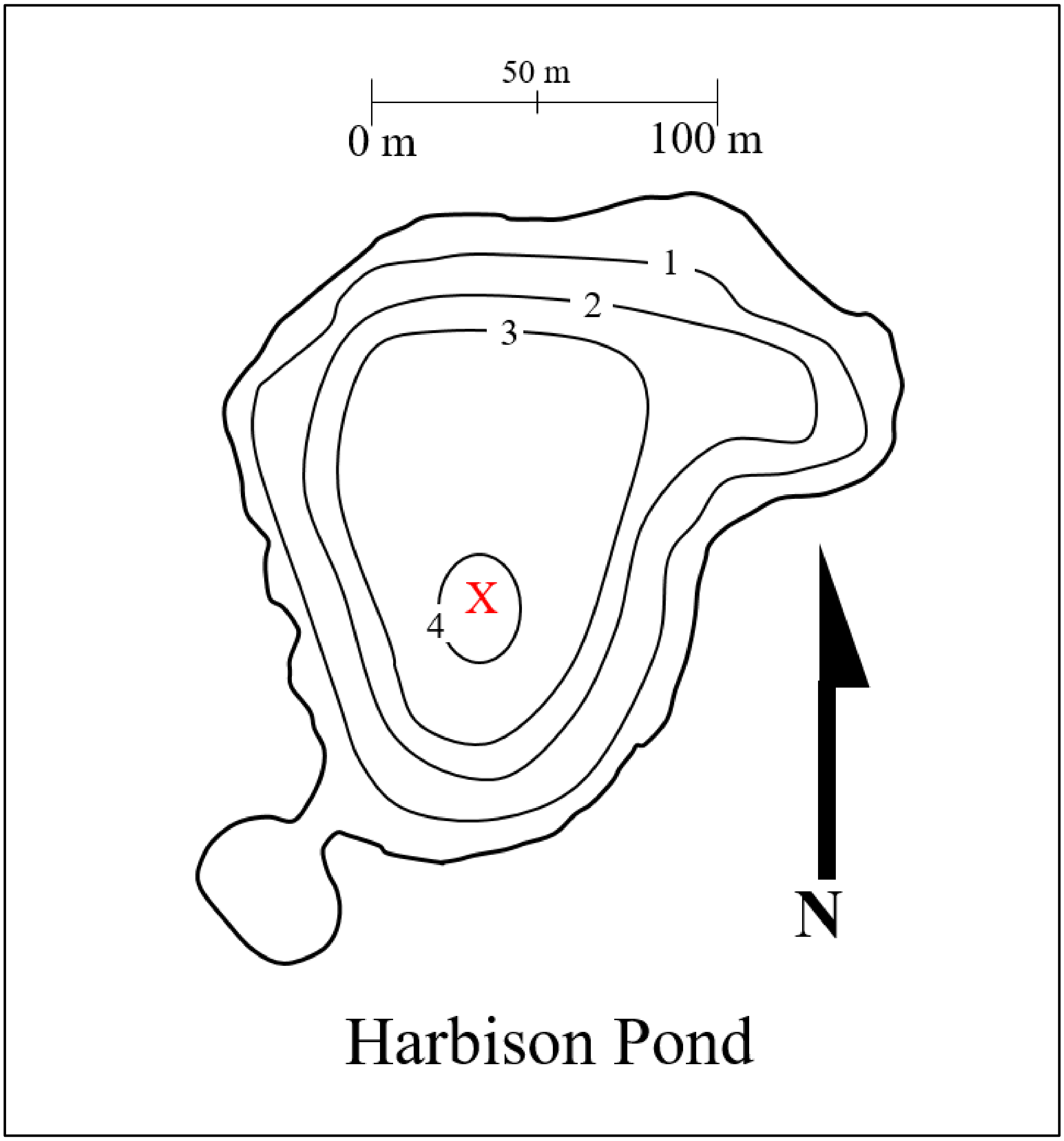

Figure 4.1. Water depth (m) of Harbison Pond.

4.3 Peat Coring

Whiskey, Pancake, and Hatchet Fens were cored at the locations where the greatest probe depths were observed using a modified Livingstone square-rod piston corer (Figure 4.3). A 0.74 m core was recovered from Pancake Fen. Coring was stopped when the corer reached unyielding rocky substrate. A 1.3 m core was recovered from Hatchet Fen. Coring was stopped when an unyielding gravel substrate was reached. Whiskey Fen yielded a 3.12 m core, with each meter extruded, wrapped, and labeled. Coring was stopped when an unyielding rocky substrate was reached. The cores were placed in protective tubes and brought to the University of Denver Sediment Laboratory.

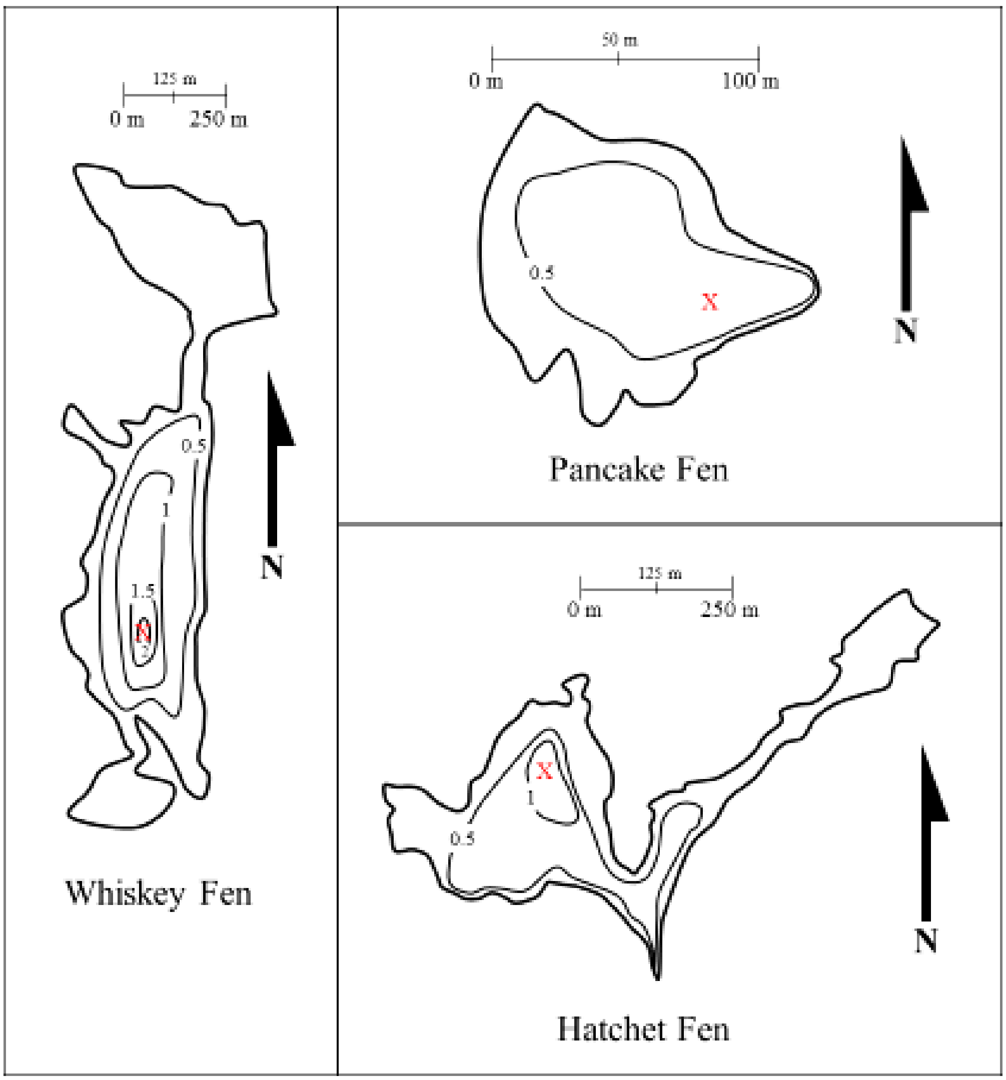

Figure 4.2. Probe depth (m) of Cored Fens.

4.4 Surface Pollen

Modern pollen rain was sampled from moss polsters on an elevational transect in and around Middle Park (Fall, 1992) (Figure 4.4). Approximately 3 cm^3 of moss material was cut and removed from polsters. The immediate surrounding vegetation was recorded. The GPS coordinates and elevation of each location was recorded with a Garmin Etrex 20x GPS. Sampling locations were at approximately 100 m elevation intervals above and below Harbison Pond on the west side of Middle Park. The samples were taken at 3,023 m, 2,928 m, 2,840 m, 2,740 m, 2,651 m, 2,560 m, 2,450 m, and 2,380 m elevation. The samples were each placed in labeled plastic bags and brought to the University of Denver Pollen Lab.

The interpretation of the pollen record recovered from Harbison Pond takes the change in modern pollen rain across an elevational transect into account when considering possible regional and local changes in vegetation. I expect the assemblage of pollen taxa within the modern samples to vary along the elevational transect, with the dominant vegetation of each elevation having increased representation in the pollen sample but *Pinus* pollen remaining dominant throughout. A great amount of variation would indicate a more local pollen signal, while very little variation would indicate a broader regional signal.

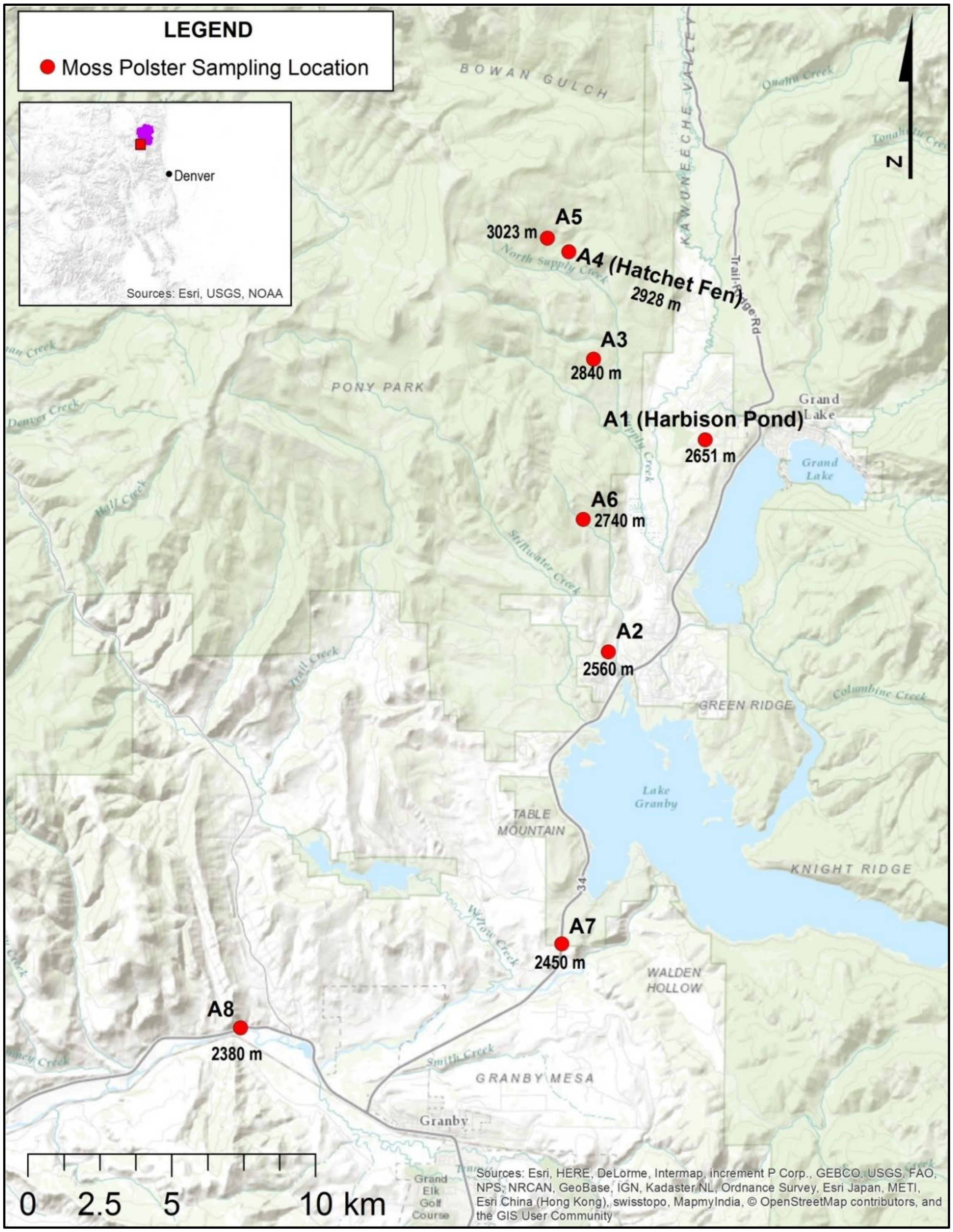

Figure 4.3. Moss polster sampling locations. Samples were taken in approximately 100 m elevation intervals at least 10 meters from any road in October, 2015.

4.5 Laboratory Methods

Peat Core

The peat core from Whiskey Fen was split and the stratigraphy of the Whiskey Fen core was logged. Colors were described while wet, using Munsell soil colors (Munsell Color, 1994).

Chronology

Four bulk sediment samples were removed for radiocarbon dating, at 190 cm at the base of the peat, 155 cm, 105 cm, and 75 cm.

Bulk Density, Organic Content, and Humification

Bulk density is a measurement of the dried mass of a given volume. Bulk density can increase with higher mineral content and decrease with higher organic content. Peat is characterized as having at least 30% organic material (Joosten and Clarke, 2002). Measuring organic content is necessary to discern between peat and peat-like material. The level of decomposition within a peat sample is measured through humification analysis. Decomposing peat produces humic acid. This humic acid is extracted during the sample preparation for humification analysis, and reduces the light transmissivity of the sample. Sampled levels that are not at least 40% organic material are not considered peat and their transmissivity values do not correspond to decomposition since there is little organic matter present to generate humic acids.

The peaty portion of the core, from 0 to 190cm, was subsampled for bulk density and organic content (Dean, 1974), and humification analysis at a 1 cm interval (Appendix C). Each 3 cm^3 sample was dried in an oven at about 95°C and a 0.2 g subsample ground into a powder with an agate mortar and pestle. Sodium hydroxide solution is added to the ground sample and simmered for one hour, and filtered. A spectrophotometer was used to measure the transmissivity of the filtered solution. Transmissivity results were corrected by dividing the transmittance by the percent organic matter. Samples with low organic matter have a corrected transmissivity greater than 100%.

Carbon Storage Calculation

Carbon storage in Whiskey Fen core was calculated, and compared to calculated storage based on estimates and techniques used by researchers with the Environmental Protection Agency (Fennessy and Nahlik, 2016). The researchers created a national carbon storage inventory, based sediment and peat samples from thousands of wetlands across the United States. This inventory provides an expected carbon storage mass for a wetland sediment depth interval of a specific classification or region. To calculate carbon storage, I used the following formula:

$$C_s = \frac{(d_1 \text{ x } (A \text{ x } 100) \text{ x } BD \text{ x } OC)}{1{,}000{,}000}$$

where C_s is carbon storage expressed in metric tons, d_1 is the depth of the peat section measured in cm, A is the area of the fen with the depth interval being calculated measured in m^2, BD is the bulk density of the peat measured in g/cm^3, and OC is the organic content of the section, expressed as a percentage. I compared this to the estimated carbon stocks of wetlands in the western United States (Fennessy and Nahlik, 2016).

Carbon storage was calculated for five different depth intervals using the deepest recoverable core depth or 120 cm, following Fennessy and Nahlik (2016). Depth intervals were used due to the variability in depth of peat and sediments in wetlands, and also within a single wetland. Using area and depth intervals rather than volume allows for direct comparison of carbon storage between wetlands of different types.

Lake Core

Sediment cores from Harbison Pond were split and the stratigraphy logged. Sediment colors were described while wet, using Munsell soil colors (Munsell Color, 1994). Samples for pollen and microscopic charcoal analysis were taken from each 2-cm interval sample of the mud-water interface, and at 2.5 cm intervals from the core.

Chronology

Nine samples were taken for radiocarbon dating. No discrete organic material was identified except those originating from aquatic plants, so all samples were bulk sediment. Samples were taken at 415 cm, 350 cm, 300 cm, 255 cm, 202 cm, 150 cm, 100 cm, 80 cm, and 42.5 cm. Samples were dated at DirectAMS in Bothell, Washington.

Pollen samples

Moss polster samples were treated with the same procedure as the sediment samples taken for pollen analysis (Appendix A). Samples for pollen processing were taken from the center of the split core to avoid contaminating with modern pollen. 1 cm^3 sample was placed in a clean plastic test tube, along with one Lycopodium tablet (Stockmarr, 1971), before being processed using standard palynological techniques (Appendix A; Berglund and Ralska-Jasiewiczowa, 1986). Processed samples were stained and mounted on slides with silicone oil. A total of 80 samples were counted at 400x magnification with a Zeiss light microscope. Samples were counted to a sum of at least 300 pollen grains per level, not including aquatic taxa. Taxa were identified based on reference materials at the University of Denver Pollen Laboratory, and published pollen keys (McAndrews *et al*., 1973; Faegri and Iverson, 1975). Pollen totals were converted to a percentage of the total counted sum (excluding aquatic taxa), and graphed in stratigraphic order. Increases or declines in the presence of pollen taxa indicate an increase or decline in abundance of the corresponding plant in the surrounding environment or region. Two pollen taxa associated with different environmental conditions can be examined as a ratio throughout a record to look for changes in relative abundance that might otherwise be difficult to detect if there is a great disparity in the relative abundance of each of them. *Pinus* and Poaceae taxa were examined as evidence of aridity and changes in forest cover in the area over time.

Microscopic Charcoal

An increase in microscopic charcoal concentration within the sediment record signifies an increase in the amount of fire within the region. A decrease in microscopic charcoal within the sediment record indicates a shift towards conditions less favorable for fire. A single sample level may represent tens to hundreds of years, so the microscopic charcoal record provides information about changes in factors that affect fire frequency within the region, such as moisture or fuel availability, but not information about specific events or episodes.

Separately from pollen counting, microscopic charcoal greater than 5 μm in diameter and *Lycopodium* spores were counted to a total sum of 200 charcoal fragments per level (Finsinger and Tinner, 2005). The *Lycopodium* control tablets that were placed with the sediment samples during processing allowed for the concentration of pollen and microscopic charcoal to be calculated. Concentration was calculated by multiplying the total number of pollen grains (or microscopic charcoal) counted by the ratio of Lycopodium spores in a tablet to the number of Lycopodium spores counted (Stockmarr, 1971).

Macroscopic Charcoal

Separate samples for macroscopic charcoal analysis (Whitlock and Millspaugh, 1996; Appendix B) were taken at 2.5 cm intervals. A syringe with the plunger removed was used to measure 3 cm^3 of sediment per level, using the plunger to gently pack the material down and remove any air pockets. The sediment samples are then broken up in a hexametaphosphate solution and bleached before being sieved. The number of visible macroscopic charcoal particles were then counted along transect lines under a macroscope at 25x magnification. Macroscopic charcoal data were analyzed using the program CharAnalysis (Higuera *et al*., 2009). CharAnalysis separates signals from noise using a threshold that is defined by the components of the charcoal record. Like with microscopic charcoal and pollen sampling, a single sample level may represent tens to hundreds of years. Also, macroscopic charcoal may continue to wash into the pond through overland flow and erosion well after the fire in which it was formed had occurred. It is also possible that isolated events that do not result in a forest fire, such as a lightning strike that does not result in a larger fire, may be a source of macroscopic charcoal within the basin that does not relate to a significant disturbance. The combination of these three factors allows for some background level of macroscopic charcoal deposition within the lake that can be regarded as "noise," which can obscure significant fire events ("signals").

Organic Content, Bulk Density and Magnetic Susceptibility

Organic content is a measurement of the mass of combustible material within a given volume of sediment. Bulk density is a measurement of the mass of a given volume of sediment. Both were determined using the procedure outlined by Dean (1974). The organic content and bulk density of lake sediment can act as proxies for temperature in Rocky Mountain lakes (D. Sullivan, personal communication). Warmer temperatures increase the amount of plant matter produced within the lake, contributing more organic matter as that plant matter decomposes. Cooler temperatures decrease the amount of organic material being produced within the lake, leading to higher bulk density as mineral-rich sediment washed into the lake makes up a greater proportion of accumulating sediment. Magnetic susceptibility is the degree of magnetization of a material in response to a magnetic field. In lake sediments, magnetic susceptibility can be used to examine changes in sediment influx over time. Sediment originating from eroded parent material containing iron and other magnetic minerals has high magnetic susceptibility, while sediment originating from plant material growing in the lake has low magnetic susceptibility.

Samples were taken from the sediment core at 5 cm intervals for measuring magnetic susceptibility. A longer sampling interval was necessary due to the 10 cm^3 needed to fill the sample pot. The sediment-water interface portion of the core had insufficient material for measuring magnetic susceptibility, and were instead sampled at a 2-cm interval for LOI and bulk density determination. Sediment samples for magnetic susceptibility were also used for measuring loss-on-ignition (LOI), and bulk density analysis. Sediment samples were first packed in 10 cm^3 plastic pots for magnetic susceptibility measurement using a Bartington MS2 magnetic susceptibility system. Due to varying degrees of moisture in the core, all pots were rinsed with deionized water before packing sediment, and drier, less malleable samples were further wetted. This was necessary to achieve accurate measurements throughout the core. After measuring magnetic susceptibility, approximately 3 cm^3 of each sample were transferred to numbered and weighed crucibles. The samples and crucibles were then reweighed, and placed in an oven at 90°C for four hours, until completely dry. Samples and crucibles were removed and allowed to cool in a desiccator for 15 minutes before again being weighed. Finally, the samples were placed in a furnace preheated to 550°C for two hours. Samples were removed from the furnace and allowed to cool for 30 minutes before again being weighed.

5. RESULTS

Whiskey Fen

Core Description

A stratigraphic diagram of the core is presented below in Figure 5.1. The basal sediments of the core are brown fine sand with some organic material from 312 cm to 292 cm depth. The sediment gradually transitions to medium coarse sand with very little organic material from 292 cm to 280 cm depth. This sand continues from 280 cm to 246 cm, before abruptly transitioning to fine granitic pebbles within a matrix of fine sand from 246 cm to 190 cm. At 190 cm depth in the core there is an abrupt transition to fibric, dark brown peat that is visually homogenous throughout the remainder of the core.

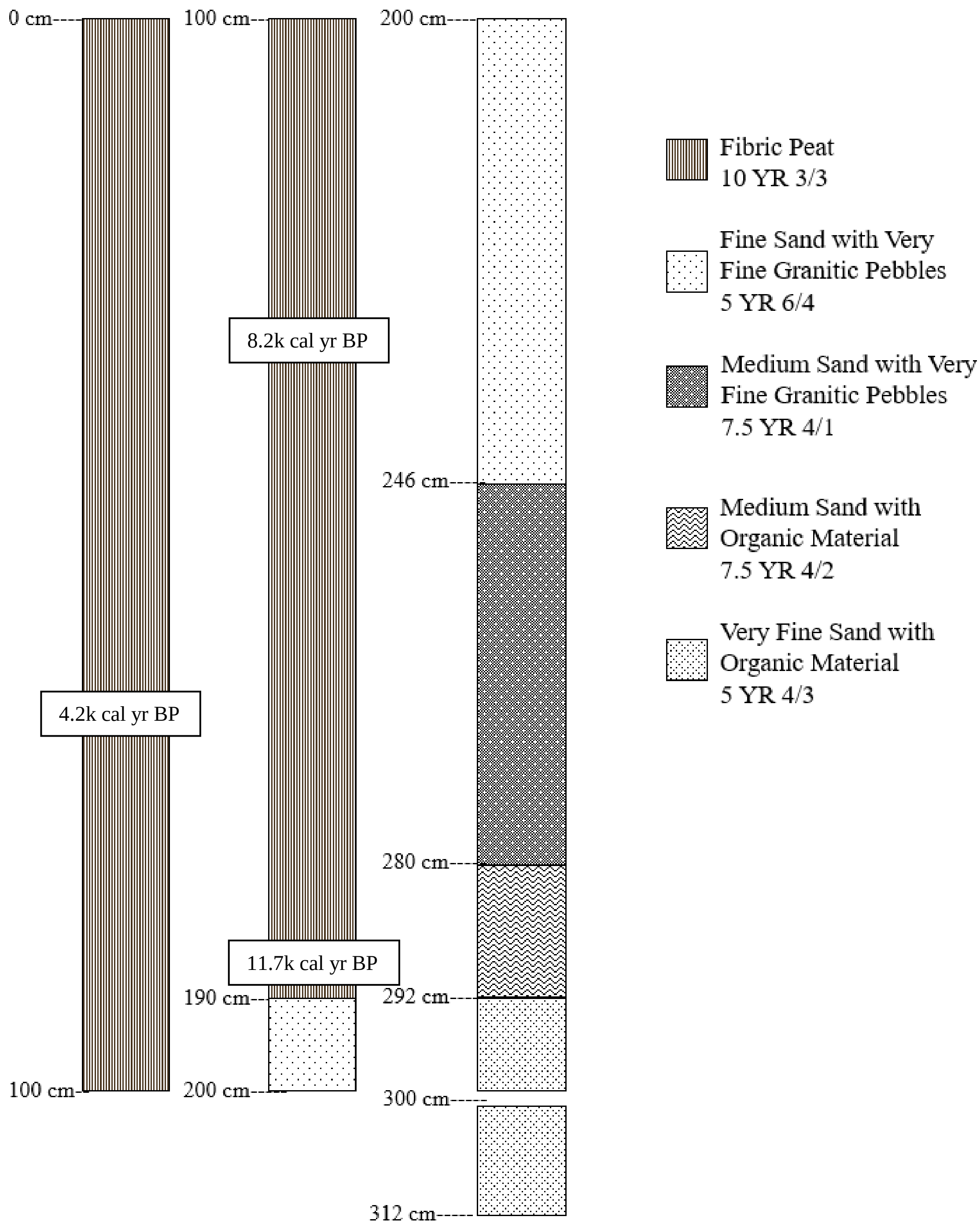

Figure 5.1. Whiskey Fen core log. Colors are described using Munsell soil colors (1994).

Chronology

Bulk sediment radiocarbon dates of the Whiskey Fen core reveal that the record extends through the Holocene (Table 5.1). A bulk peat sample taken from the base of the peat section yielded a radiocarbon date of 10,512 C^{14} yrs BP. A peat sample from 155 cm yielded a radiocarbon date of 8,579 C^{14} yrs BP. A peat sample from 105 cm yielded a radiocarbon date of 5,879 C^{14} yrs BP. A peat sample from 75 cm yielded a radiocarbon date of 2,063 C^{14} yrs BP. These radiocarbon dates were calibrated using Bacon and the IntCal13 carbon curve (Blaauw and Christen, 2011; Reimer *et al*., 2013). An age-depth diagram and the sedimentation rate of Whiskey Fen is presented in Figures 5.2 and 5.3.

Table 5.1. Uncalibrated and calibrated radiocarbon dates from Whiskey Fen. Radiocarbon dates were calibrated using the program Bacon (Blaauw and Christen, 2011) and the IntCal13 carbon curve (Reimer *et al*., 2013).

Sample ID	**Depth (cm)**	**Radiocarbon Age**	**cal yr BP**	**Material**
WF-75	75	2,063 ± 33	2,069	peat
WF-105	105	5,879 ± 42	6,727	peat
WF-155	155	8,579 ± 33	9,559	peat
WF-190	190	10,512 ± 45	12,347	peat

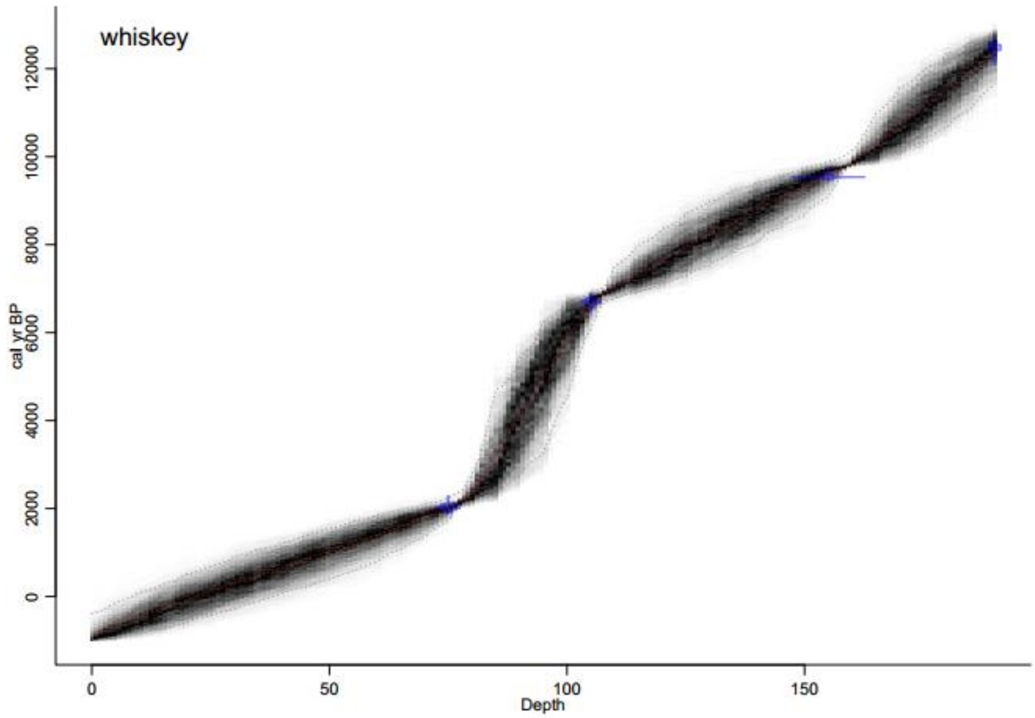

Figure 5.2. Age-depth diagram of Whiskey Fen. This diagram was generated using the program Bacon (Blaauw and Christen, 2011) and the IntCal13 carbon curve (Reimer *et al.*, 2013). Darker colored areas indicate more likely values. The red line indicates the "best fit" model for age-depth. The blue shapes indicate radiocarbon ages and their level of uncertainty.

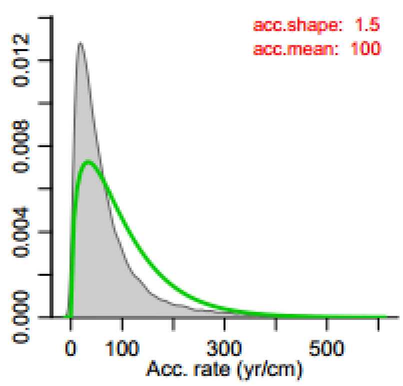

Figure 5.3. Sediment accumulation rate in Whiskey Fen. Generated using the program Bacon (Blaauw and Christen, 2011) and the IntCal13 carbon curve (Reimer *et al.*, 2013).

Bulk Density and Organic Matter

The bulk density determinations of the peat section of Whiskey Fen are shown in Figure 5.4. Bulk density declines from a peak of about 0.72 g/cm^3 at its base of 190 cm to fluctuating between about 0.15–0.40 g/cm^3 from about 105–184 cm. The bulk density of Whiskey Fen is greatest between about 74–102 cm. Bulk density steadily declines from about 0.40 g/cm^3 at 65 cm to about 0.15 g/cm^3 until the top of the core (present).

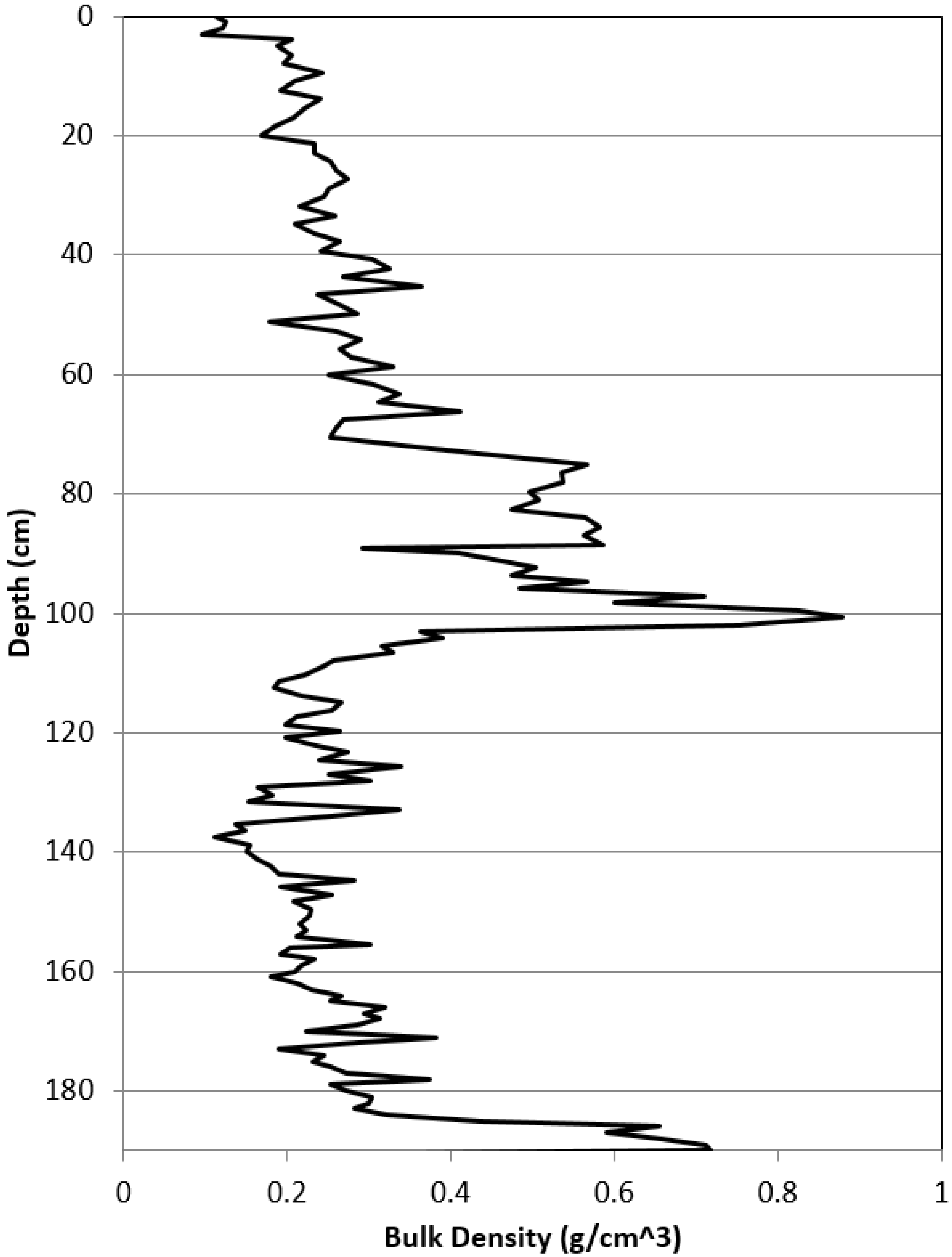

Figure 5.4. Bulk density (g/cm^3) of Whiskey Fen by depth.

Organic matter content is presented below in Figure 5.5. Organic matter content ranged from about 7% to over 80%.

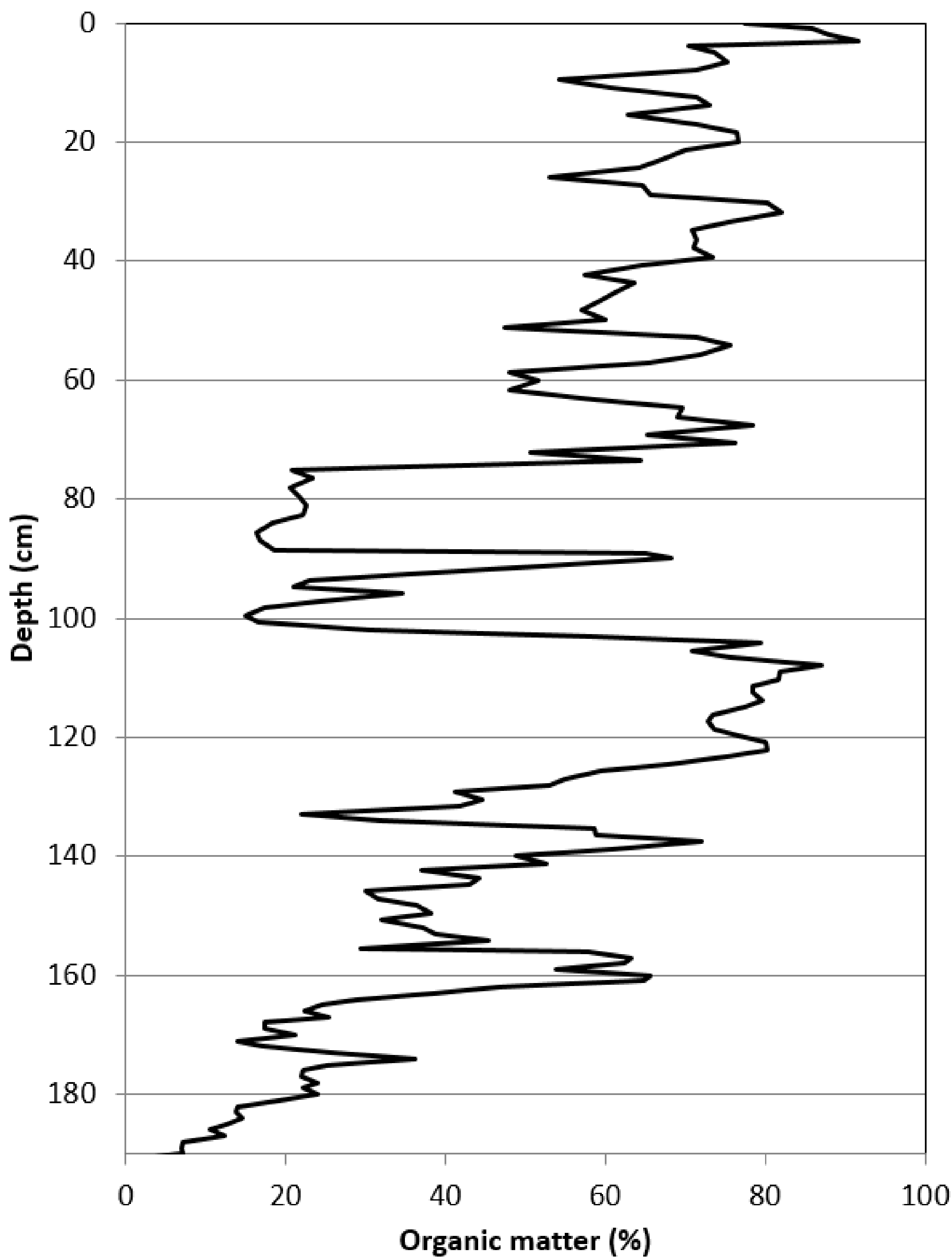

Figure 5.5. Organic matter content (%) of Whiskey Fen by depth.

Humification

Humification analysis of the peat from Whiskey Fen is presented in Figures 5.6 and 5.7. Transmissivity values over 100% are a result of correcting for organic matter. The transmissivity of a sample is divided by the organic content of the sample. Samples with low organic content values tend to have high transmissivity values. For this reason, humification values for samples with less than 40% organic matter content are frequently over 100% transmissivity after correction and indicate a lack of peat.

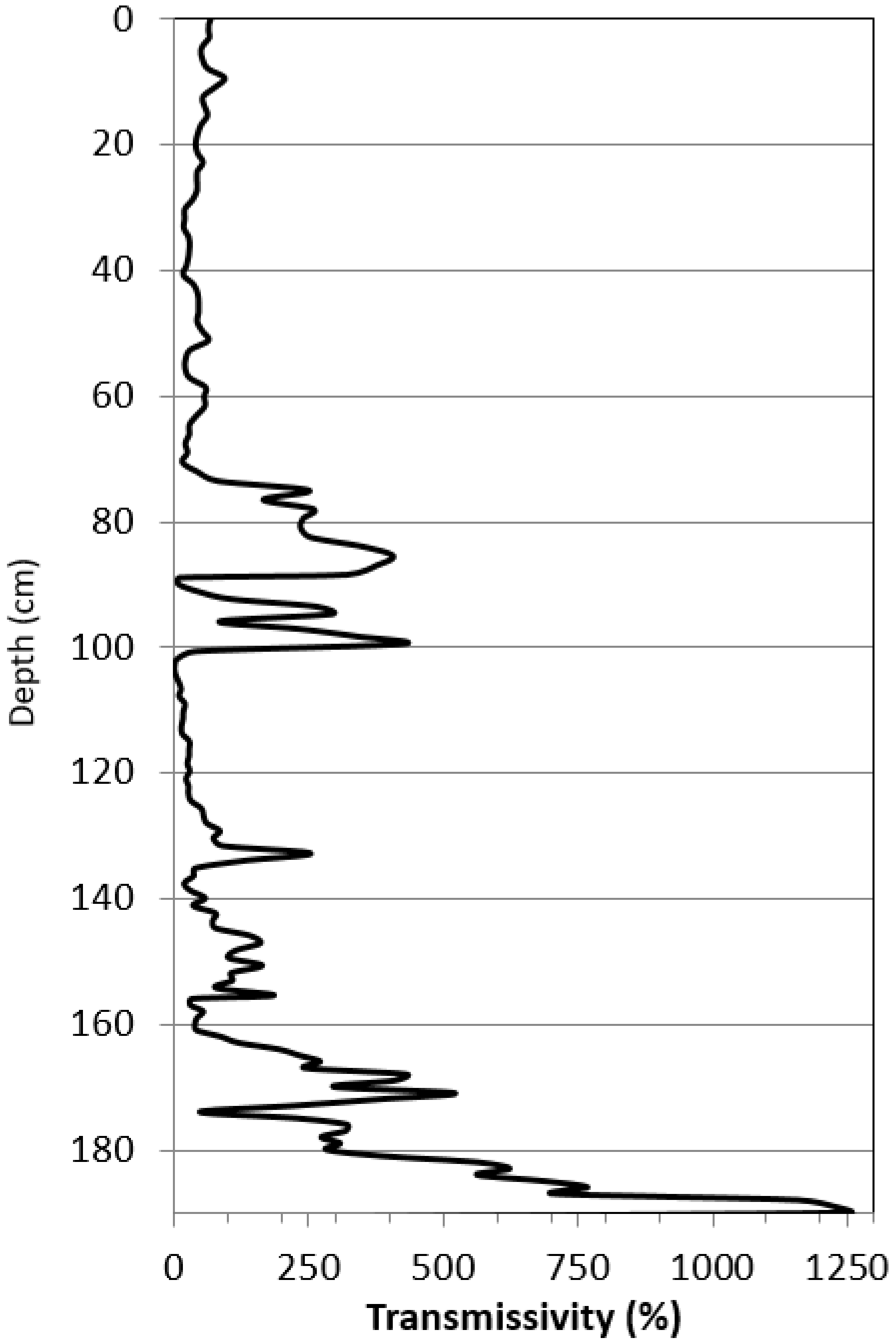

Figure 5.6. Transmissivity (%) of Whiskey Fen by depth. Samples are calibrated by organic matter content (%).

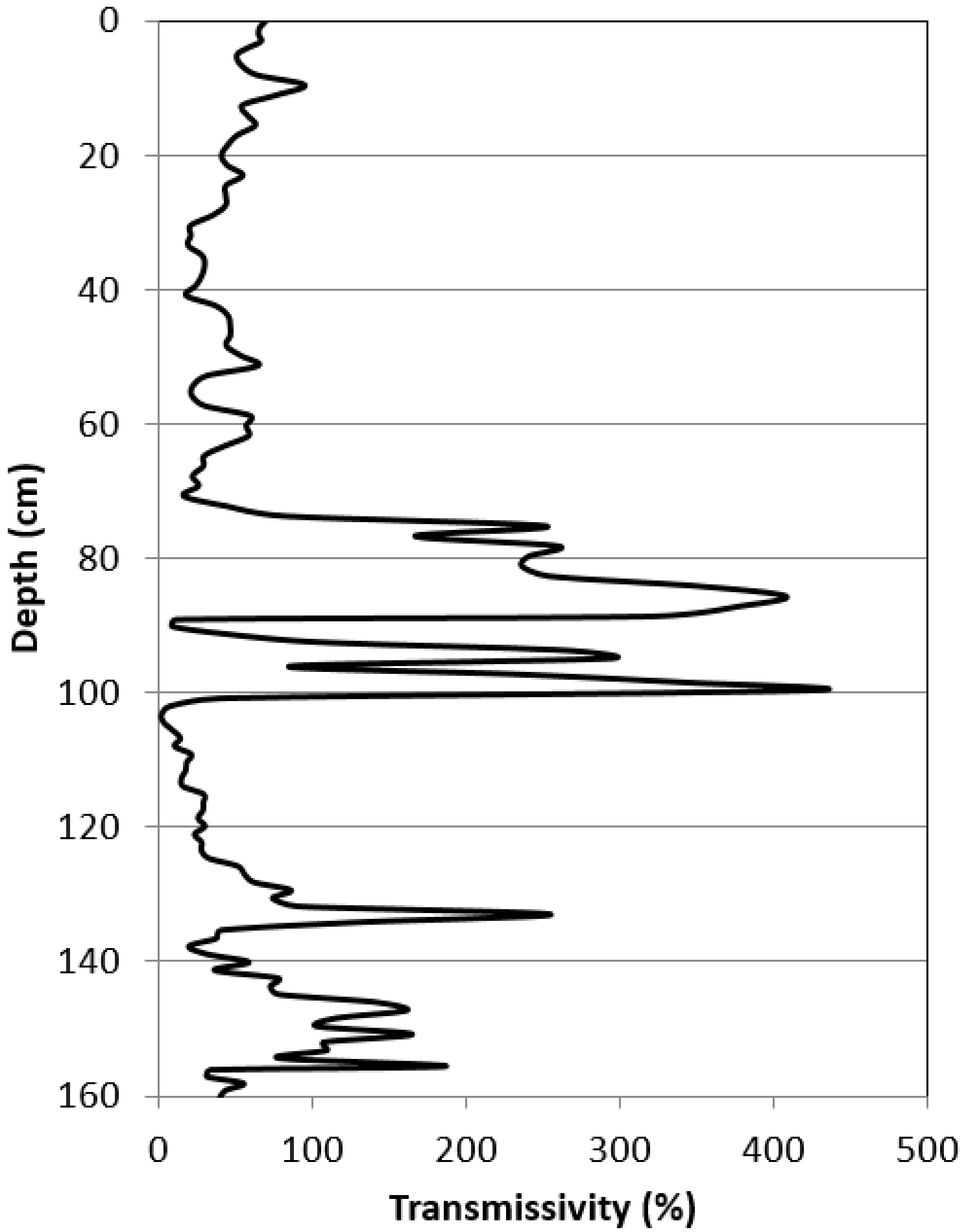

Figure 5.7. Transmissivity (%) of Whiskey Fen, 0–160 cm. Samples are calibrated by organic matter content (%).

Carbon Storage

The organic carbon storage of Whiskey fen by depth interval is presented in Table 5.2. The calculated total storage of the three surveyed fens is shown in Table 5.3.

Table 5.2. Calculated organic carbon storage (kg/m^2) by depth intervals, following Fennessy and Nahlik, 2016.

Depth Interval	**Organic Carbon Storage (kg/m^2)**
Calculated 0–10 cm	6.62
Calculated 0–30 cm	26.22
Calculated 31–60 cm	36.47
Calculated 61–90 cm	41.56
Calculated 91–120 cm	59.244
Total 120 cm core (kg/m^2)	**163.48**

Table 5.3. Total organic carbon storage by depth level.

	Hatchet Fen	**Pancake Fen**	**Whiskey Fen**
Area with less than 50 cm peat depth (assuming average depth of 25 cm)	2,309.26 tons	134.93 tons	5,895.78 tons
Area with 50 cm peat depth	854.26 tons	44.15 tons	1,312.54 tons
Area with 100 cm peat depth	523.25 tons	N/A	3,303.55 tons
Area with 150 cm peat depth	N/A	N/A	175.05 tons
Area with 200 cm peat depth	N/A	N/A	37.13 tons
Total	**3,686.77 tons**	**179.08 tons**	**10,724.05 tons**

Harbison Pond

Core Description

The Harbison Pond core log is shown in Figures 5.8 and 5.9. The bottom of the core from 418 cm to 383 cm is composed of dark grey sediment composed of laminations of fine sand and clay. From 383 cm to 221 cm, the sediment is lighter in color and the laminations are more distinct. Rocky Mountain lily seeds (*Nuphar polysepala)* are present from this section to the top of the core. An abrupt transition occurs at 162 cm, as clay and fine sand are no longer the dominant components of the sediment. The sediment becomes darker and the organic material is intermixed with clay and fine sand from 162 cm to 155 cm. A peat-like layer of mostly fibrous biological matter extends from 155 cm to 28 cm. The top 28 cm of the core is unconsolidated sediment, and presents as a dark brown liquid.

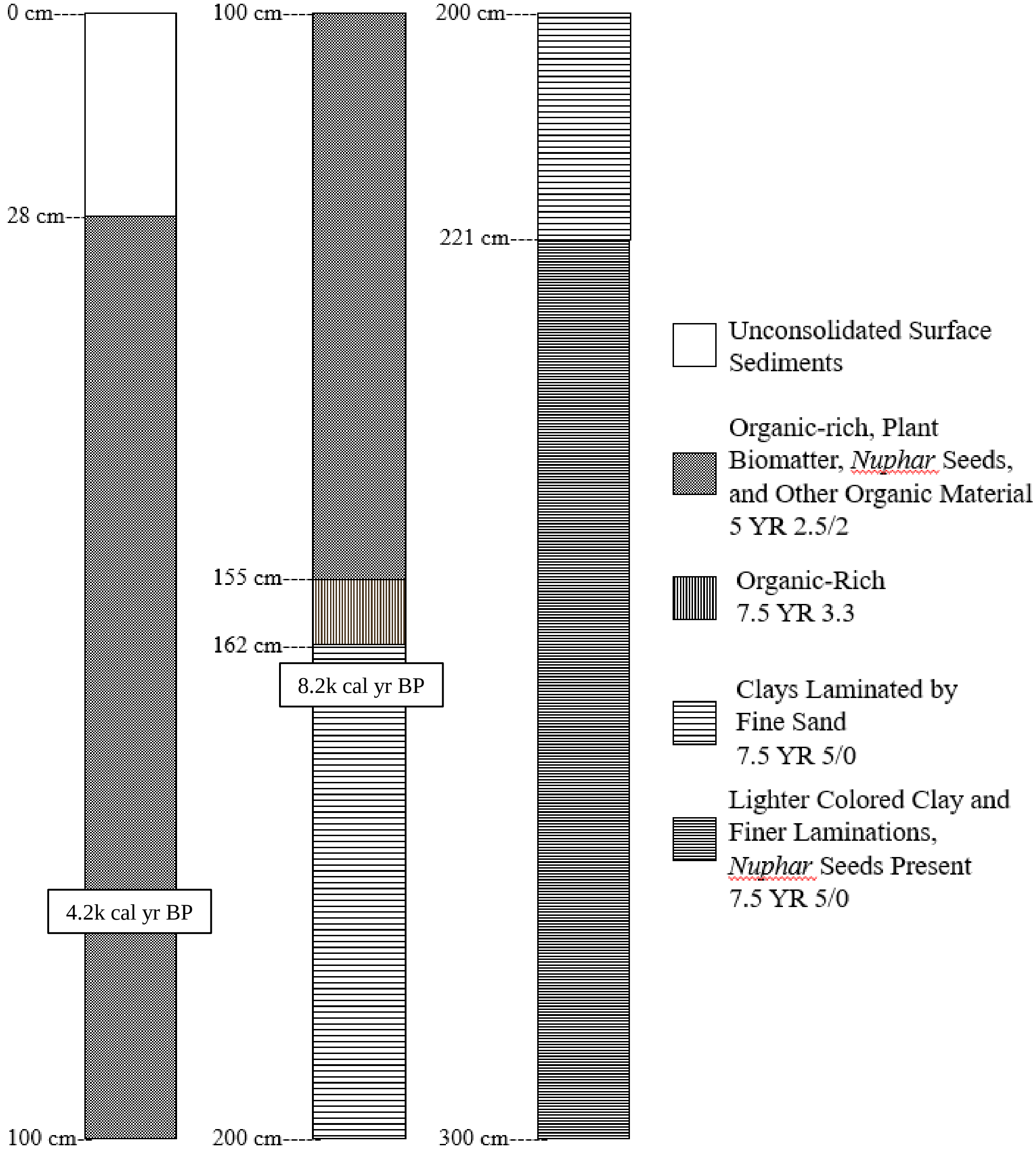

Figure 5.8. Harbison Pond core log. Colors are described according to Munsell soil colors (1994).

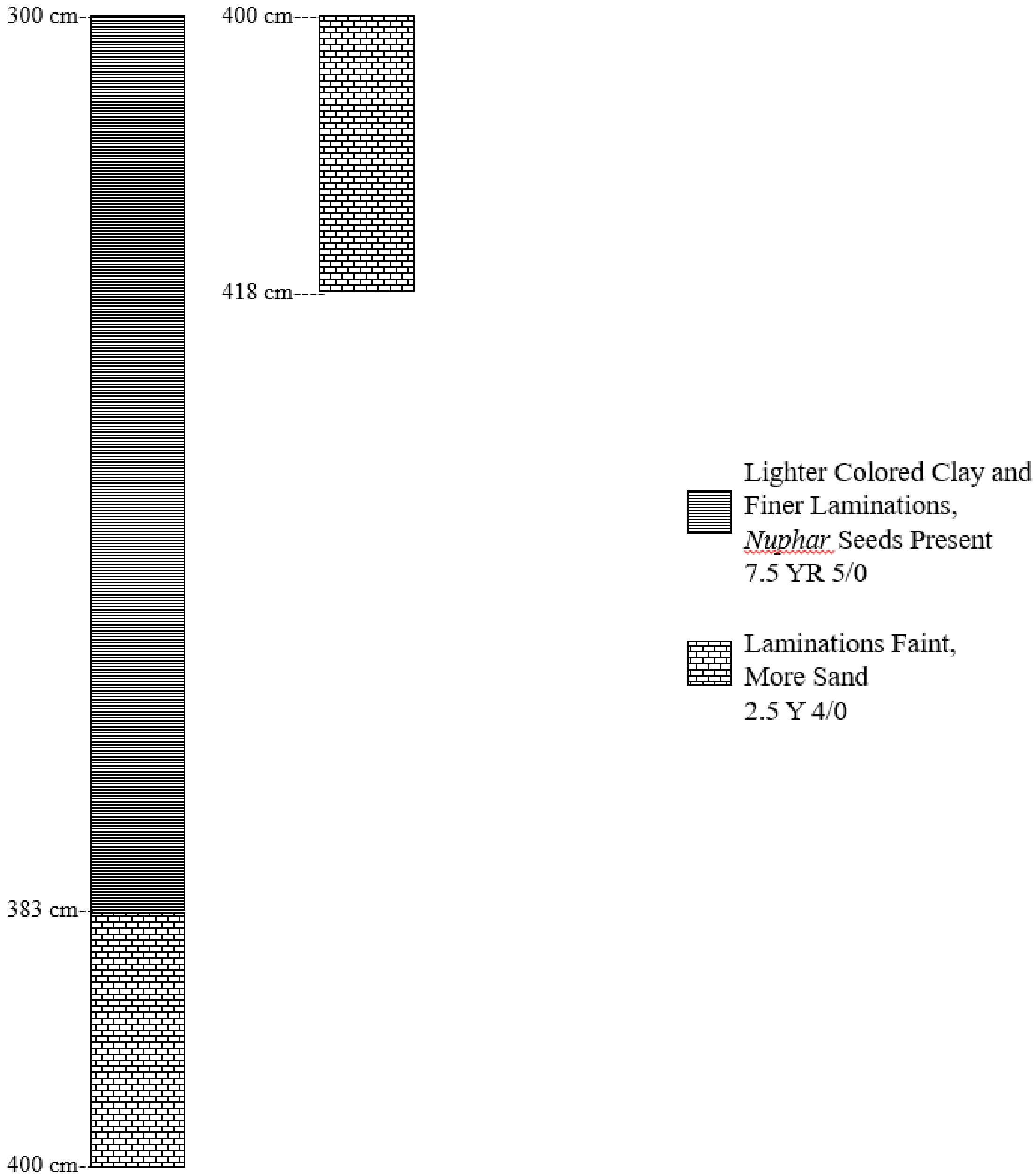

Figure 5.9. Harbison Pond core log (continued). Colors were described following a Munsell colors (1994).

Chronology

The radiocarbon ages of the dated samples are shown in Table 5.4 and the age-depth diagram and sediment accumulation rate are shown in Figures 5.10 and 5.11, respectively. Bulk sediment radiocarbon dates of the Harbison Pond core reveal that the record extends into the late Pleistocene. A sample taken from the base of the core dates to 9,905 C^{14} yr BP, but bulk sediment sampled at 202 cm and 255 cm date to 14,072 C^{14} yr BP and 14,592 C^{14} yr BP, respectively. Material from 350 cm yielded a radiocarbon age of 14,940 C^{14} yr BP. Like the basal date, the sample taken from 300 cm yielded a reversed date.

Sediment samples can be biased towards younger radiocarbon dates through increased bioturbation, the downward growth of rootlets of aquatic vegetation, or other processes that could introduce living organisms into underlying layers of sediment. Sediments can be biased towards older radiocarbon dates through the failure and release of material from shelves or steep slopes within the lake. Harbison Pond is a flat-bottomed lake and the core was taken from the center. Aquatic vegetation is covers a majority of the lake. I did not include the younger radiocarbon date when reversals occurred, since the likelihood of younger material being introduced lower in the sediment record of Harbison Pond is greater than the likelihood of older organic material being introduced higher in the sediment record.

The radiocarbon dates were calibrated using the IntCal13 carbon curve for northern hemisphere sites (Reimer *et al*., 2013) and the statistical program Bacon (Blaauw and Christen, 2011). Bacon generates an age-depth model using the provided radiocarbon dates and selected carbon curve, and subdividing the core a specified number of times to run separate statistical calculations of sedimentation rates and then fitting them together (Figure 5.1). The age-depth curve is used to calculate approximate dates used in the discussion in chapter 6.

Table 5.4 Raw and calibrated radiocarbon dates of bulk sediment samples from Harbison Pond. Radiocarbon dates were calibrated using the program Bacon (Blaauw and Christen, 2011) and the IntCal13 carbon curve (Reimer *et al*., 2013). *Samples were not used in constructing a chronology.

Sample ID	**Depth (cm)**	**Radiocarbon Age**	**cal yr BP**	**Material**
BP1-42.5	42.5	1353 ± 40	1279	gyttja
BP1-80	80	4391 ± 40	4965	gyttja
*BP1-100	100	3365 ± 29	3665	gyttja
BP2-150	150	6,772 ± 32	5,727	gyttja
BP1-202	202	14,072 ± 47	15,182	clastic
BP1-255	255	14,592 ± 70	15,823	clastic
*BP2-300	300	13,910± 86	14,892	clastic
BP2-350	350	14,940 ± 51	16,201	clastic
*BP2-415	415	9,905 ± 44	9,363	clastic

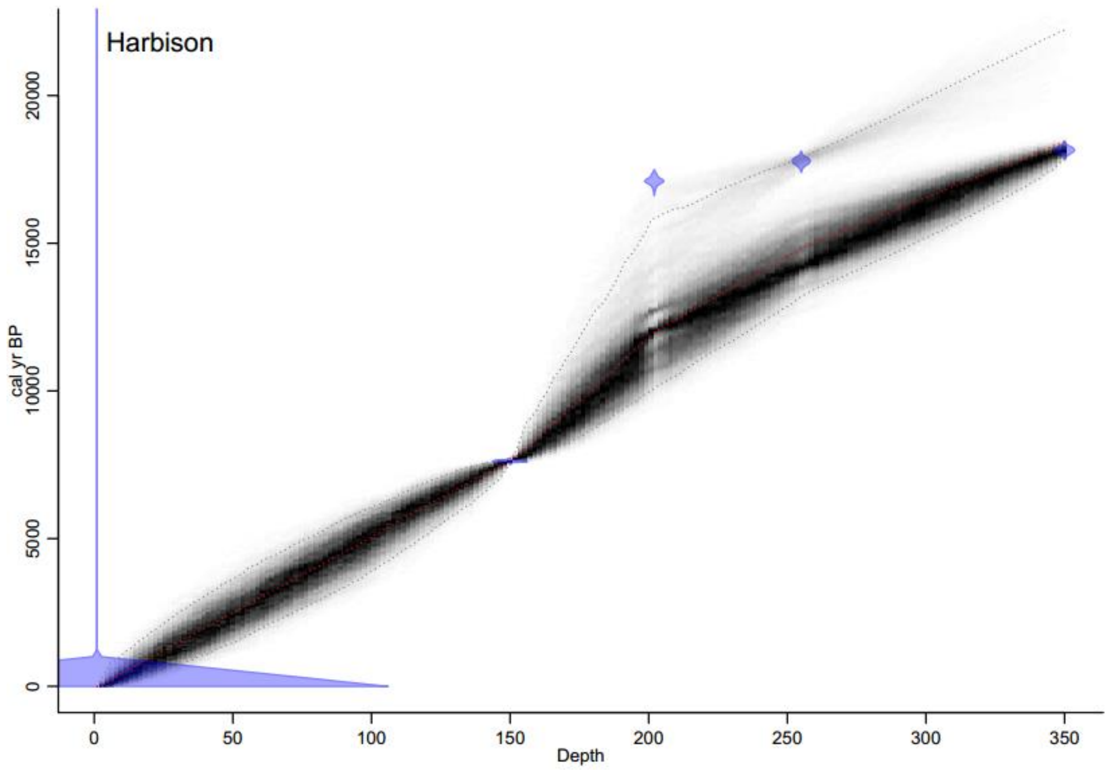

Figure 5.10. Age-depth diagram of Harbison Pond. Generated using the program Bacon (Blaauw and Christen, 2011) and the IntCal13 carbon curve (Reimer *et al*., 2013). Darker colored areas indicate more likely values. The red line indicates the “best fit” model for age-depth. The blue shapes indicate radiocarbon ages and their level of uncertainty.

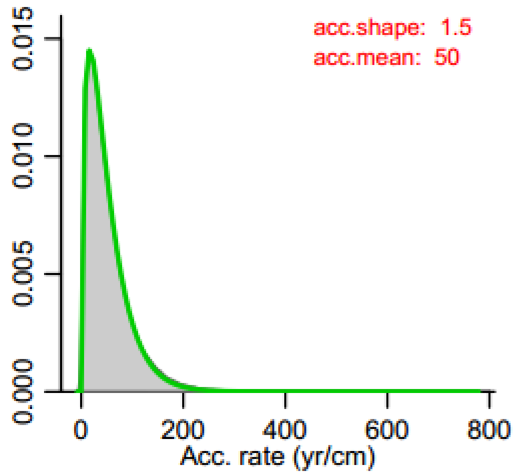

Figure 5.11. Sediment accumulation rate in Harbison Pond. Generated using the program Bacon (Blaauw and Christen, 2011) and the IntCal13 carbon curve (Reimer *et al*., 2013).

Microscopic Charcoal, Magnetic Susceptibility, Bulk Density, and Organic Content

The results of the charcoal analysis are shown in Figure 5.12. From about 9,000 yr BP to about 7,000 yr BP the microscopic charcoal record shows distinct peaks that are the highest in the record. Charcoal concentrations decline from about 8,200 to 7,800 yr BP. From 7,000 yr BP to 2,800 yr BP, the concentration of microscopic charcoal fluctuates less but remains consistently higher than from 2,800 yr until present.

Magnetic susceptibility is consistently high in the bottom portion of the core, with the exception of a decline centered at about 6,900 yr BP. Magnetic susceptibility greatly declines at about 4,250 yr BP and remains low for the rest of the record, as the organic content increases above 50%. Bulk density is presented in Figure 5.13. Bulk density is highest at the beginning of the record, and begins decreasing at 210 cm and continues to the top of the core.

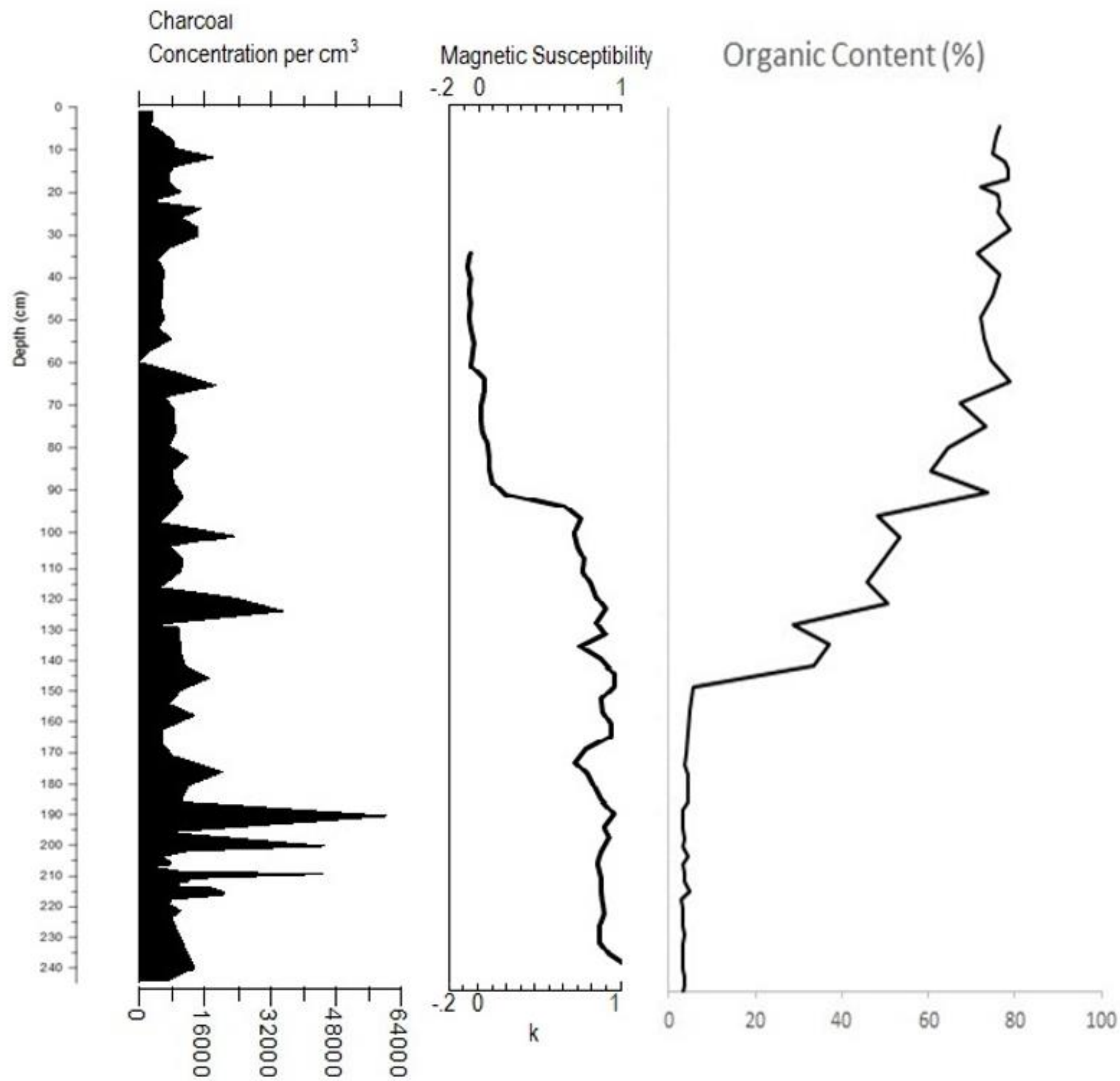

Figure 5.12. Microscopic charcoal, magnetic susceptibility, and organic content of the Harbison Pond sediment core.

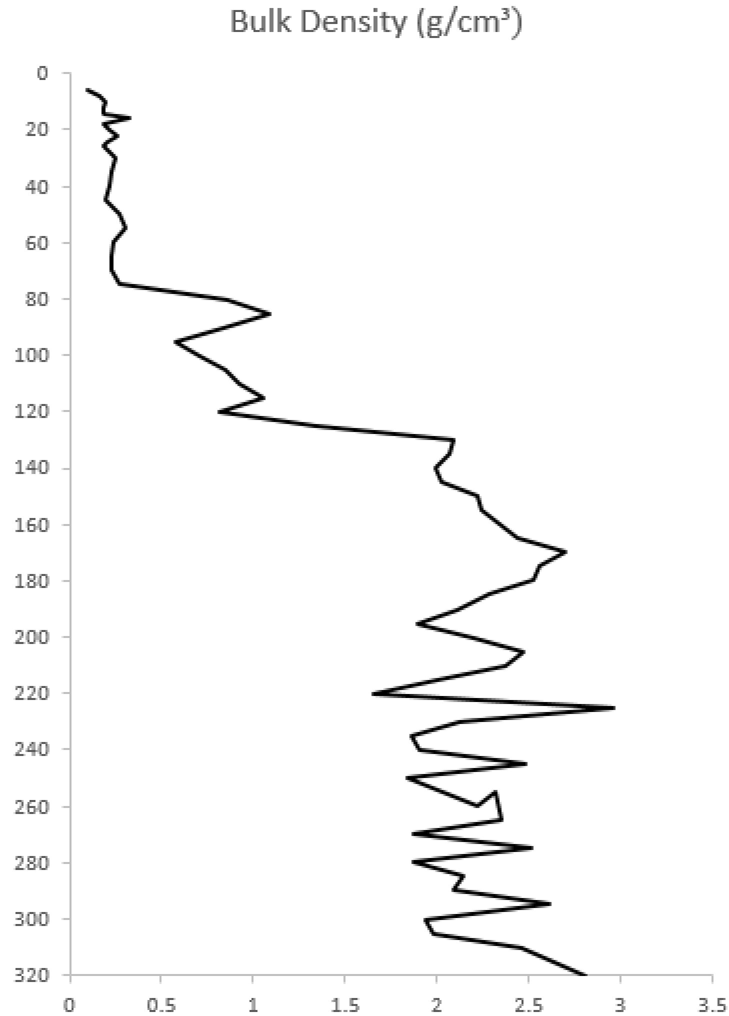

Figure 5.13. Bulk density of the Harbison Pond sediment core.

Macroscopic Charcoal

CharAnalysis is a publicly available program that that aids in analyzing charcoal records and analyzing charcoal peaks, utilizing statistical and graphical tools (Higuera *et al*., 2009). The charcoal record was converted from depth to age for the purpose of this analysis. CharAnalysis indicates there are seven major charcoal peaks in the record, with the highest concentrations occurring between 9,000 and 5,400 yr BP (Figure 5.14). FRI and fire frequency are estimated from the record. Fire frequency is observed to be greatest from about 8,500–6,000 cal yr BP.

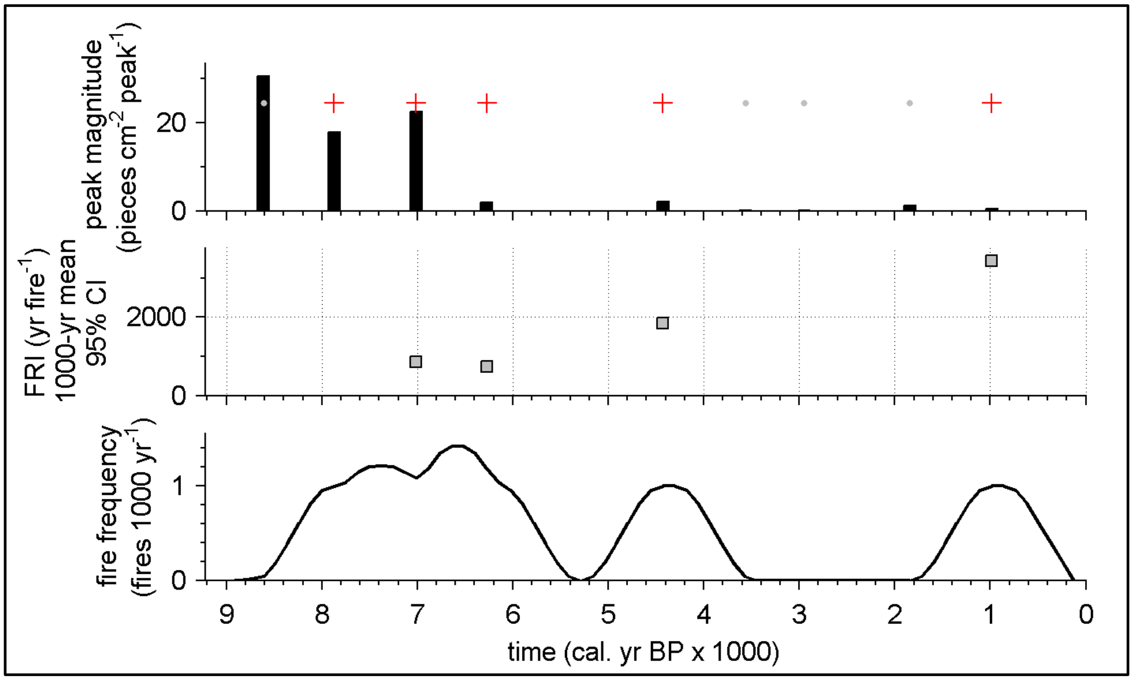

Figure 5.14. Peak magnitude, fire-return interval (FRI), and fire frequency from the macroscopic charcoal record. Peak magnitude bars measure fire size and intensity. The red "+" symbols mark fire episodes, while the grey dots indicate small peaks that do not meet statistical significance. Fire frequency shows the total number of fires in a 1000-yr period.

Surface Pollen

Modern pollen rain sampling along an elevational transect is presented in Figure 5.15.

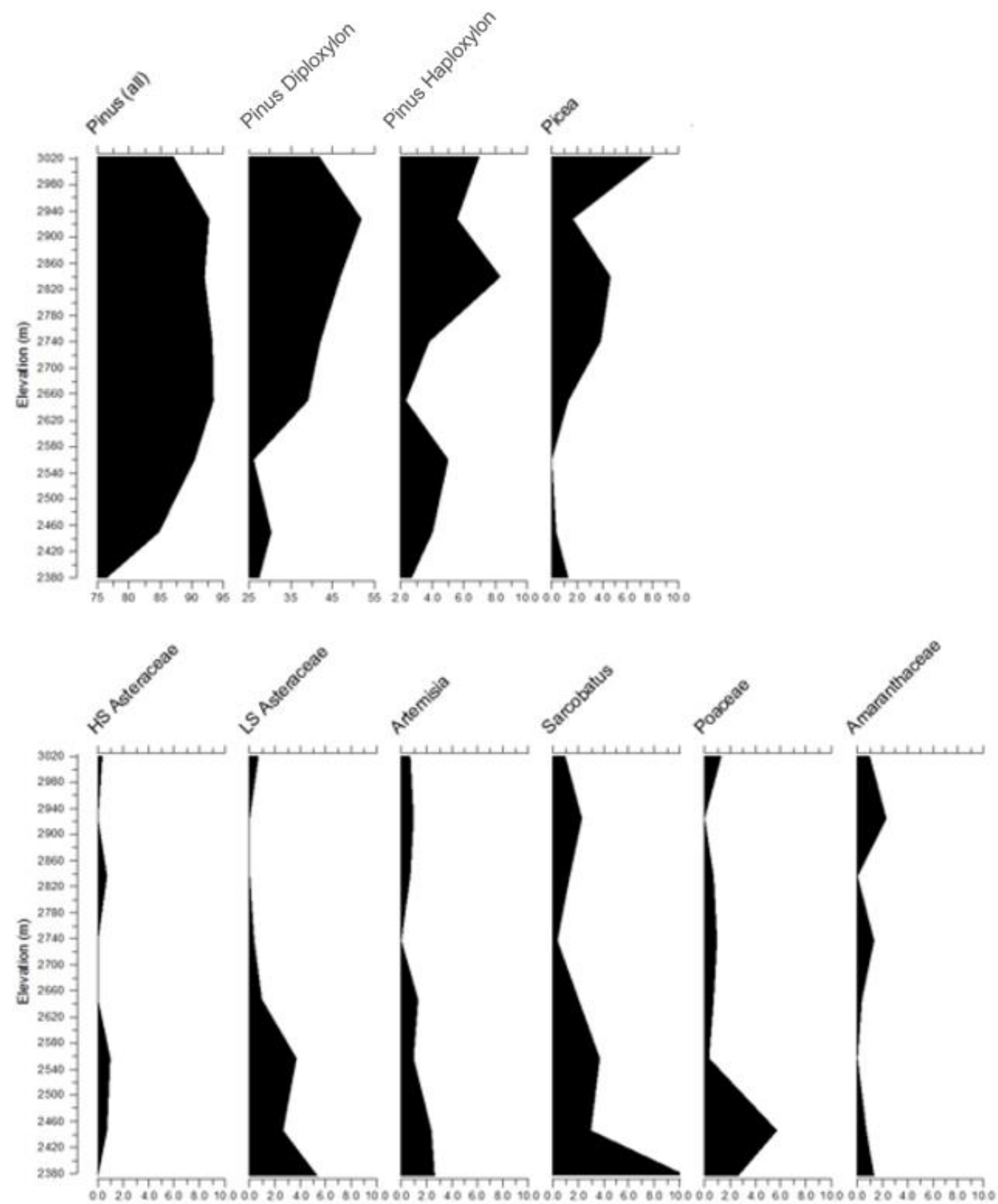

Figure 5.15. Modern pollen rain taxa of Middle Park on the elevational transect.

Pollen Analysis

Pollen diagrams from Harbison Pond are presented in figures 5.16 through 5.18 below. Figure 5.16 shows the percentages of the most important arboreal taxa, Figure 5.17 shows the most significant non-arboreal taxa, and Figure 5.18 shows the most important aquatic taxa. The pollen taxa included in the group “Other Arboreal” is a combination of all tree taxa besides *Pinus.* The pollen taxa that make up the group “Herbaceous and Grasses” in figure 5.17 is a combination of Amaranthaceae, *Artemisia*, low-spine and high spine Asteraceae, *Ambrosia*, Poaceae, and *Sarcobatus*.

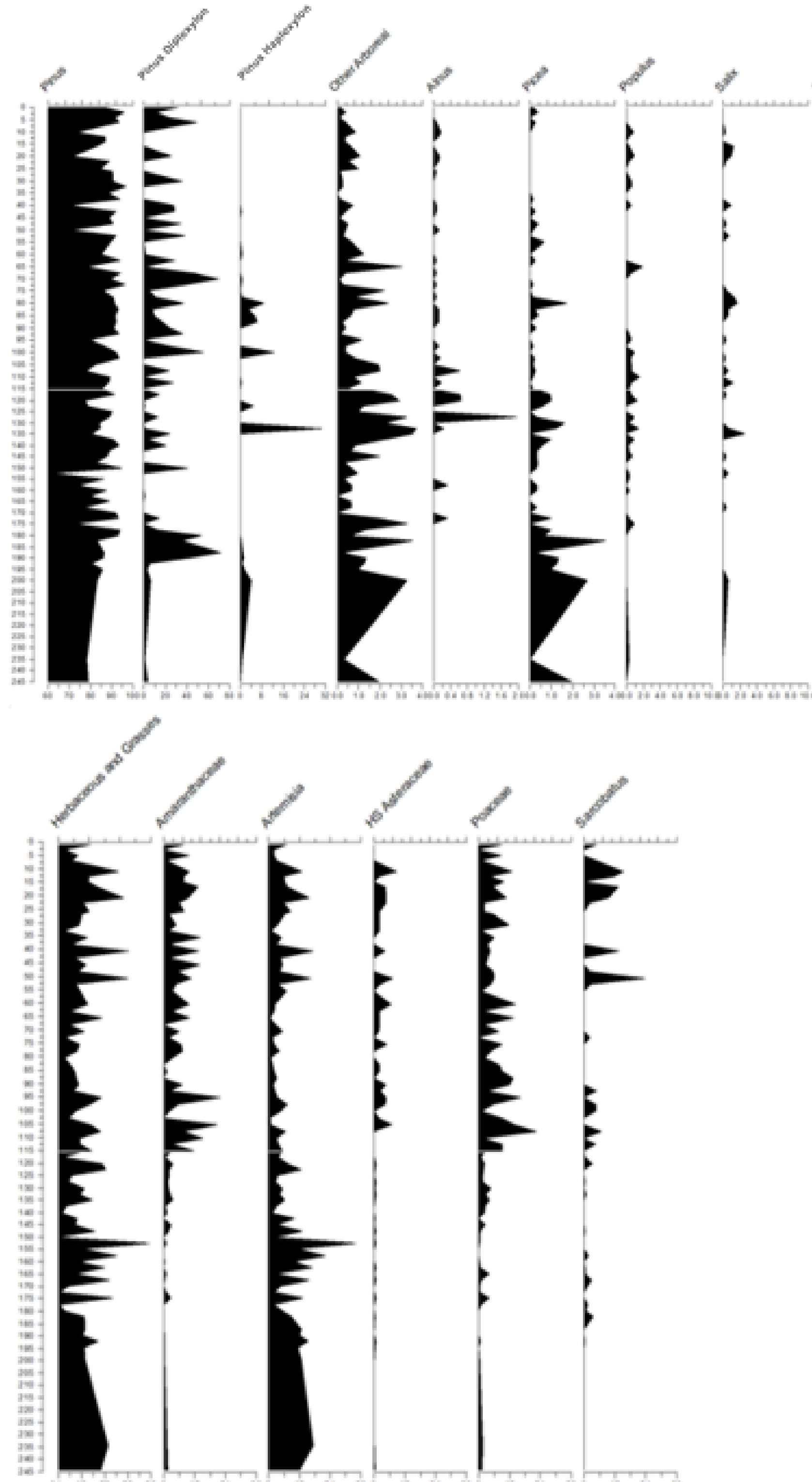

Figure 5.16. Harbison Pond pollen diagram.

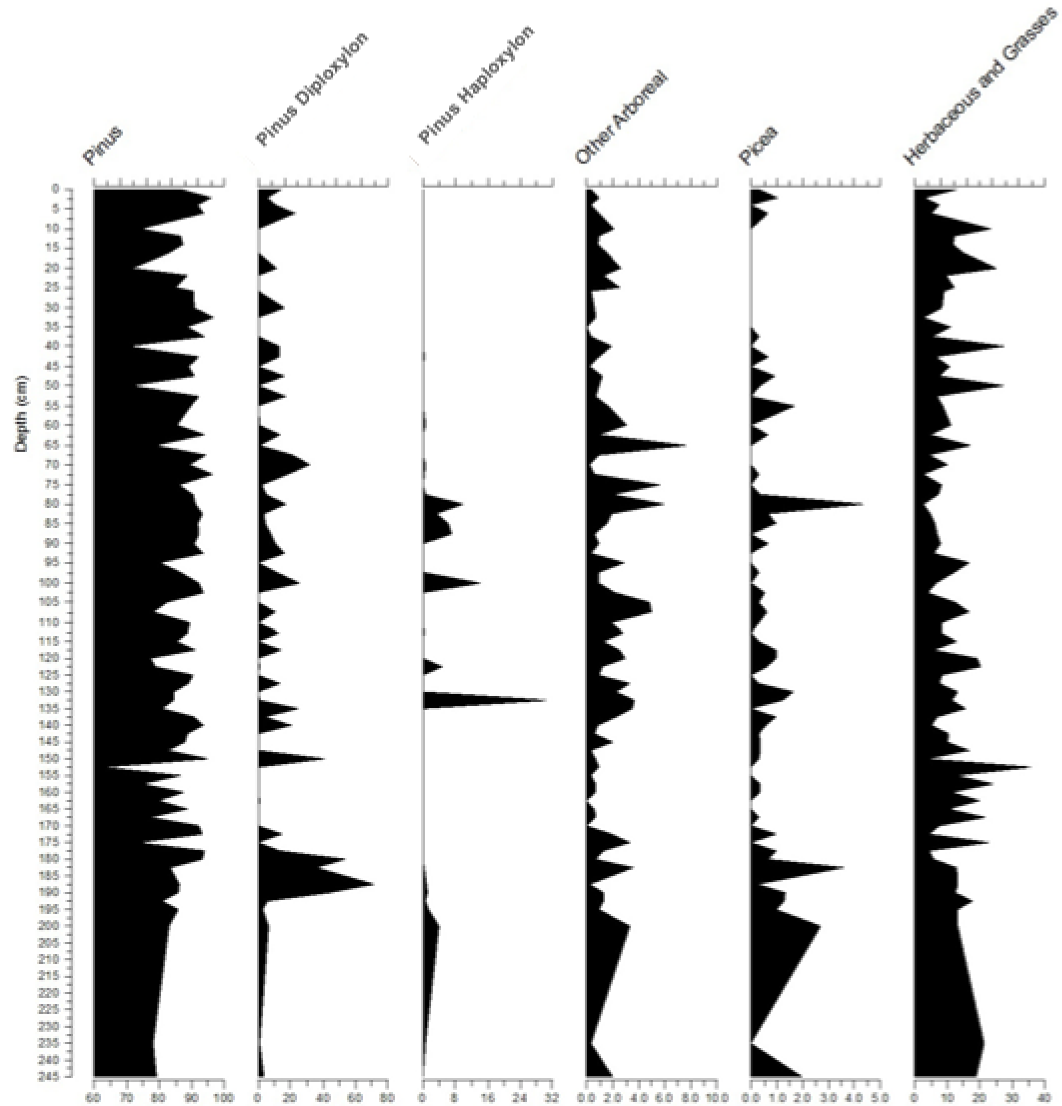

Figure 5.17. Harbison Pond summary pollen diagram. *Pinus* taxa are dominant throughout the core.

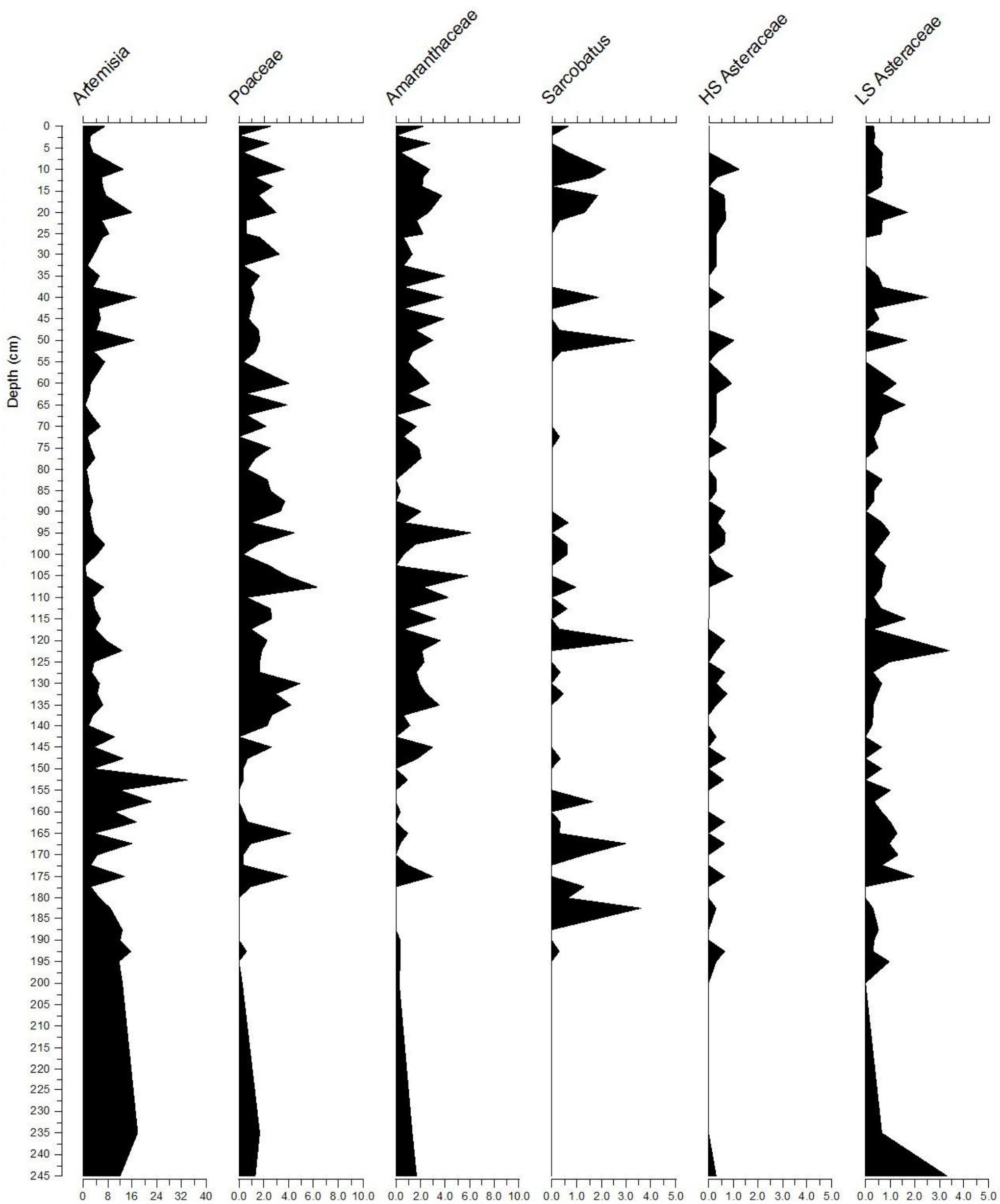

Figure 5.18. Herbaceous and grass pollen diagram from Harbison Pond.

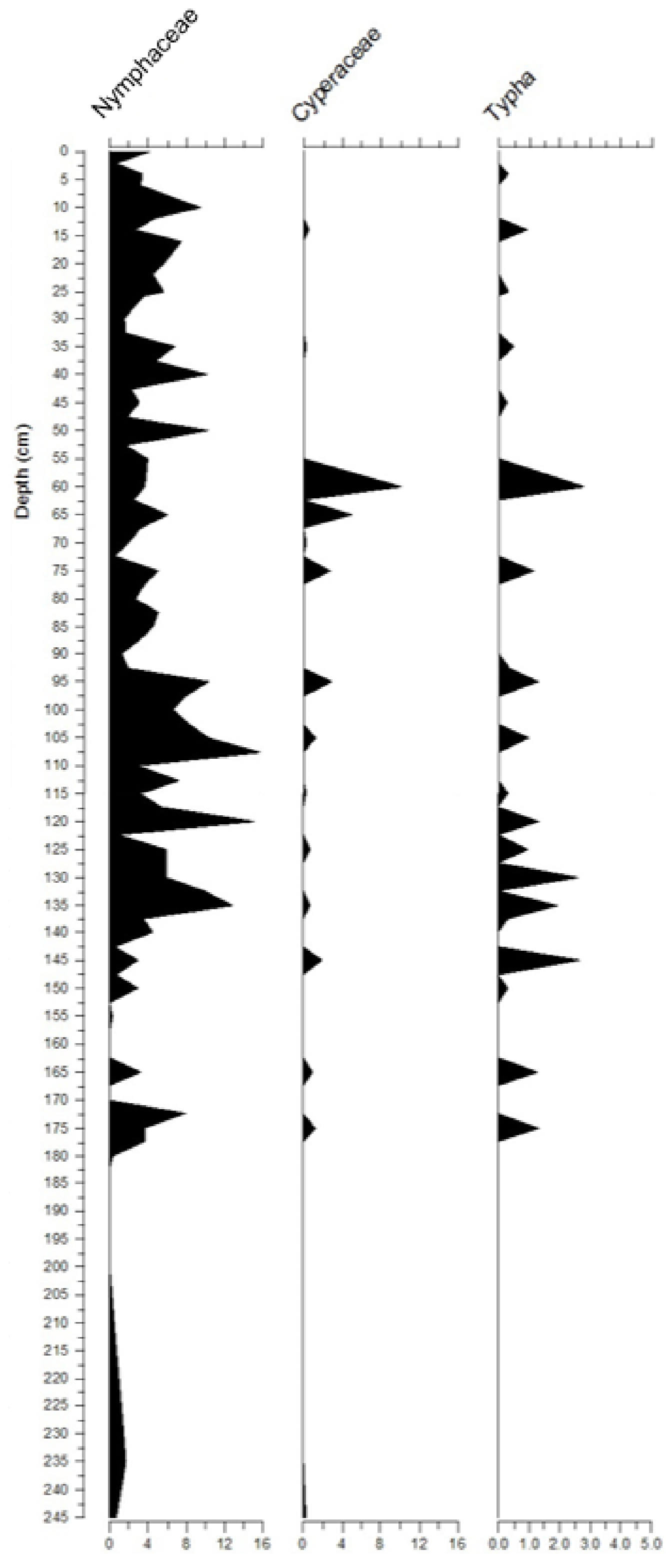

Figure 5.19. Aquatic pollen diagram from Harbison Pond.

Pollen Concentration and Poaceae/Pinus Ratio

Pollen concentration and the ratio between Poaceae and *Pinus* pollen are presented in Figure 5.20. Pollen grain concentration is too low to count below 245 cm.

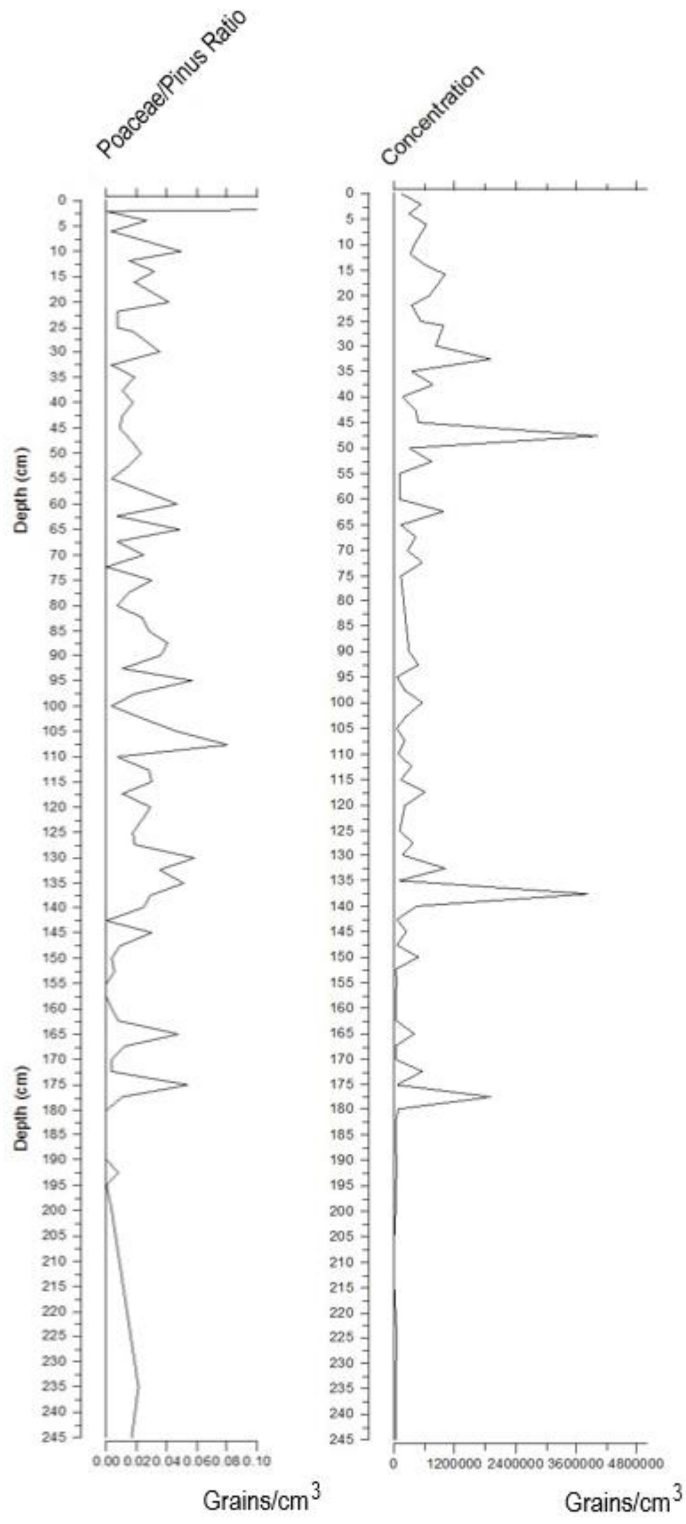

Figure 5.20. Poaceae/Pinus ratio and pollen concentration.

6. DISCUSSION

Whiskey Fen

Description and Chronology

The material in the lower portion of the core, below 190 cm, did not contain peat. The core contains granitic pebbles within a matrix of sand between 280 cm and 190 cm, indicating that prior to the formation of Whiskey fen approximately 12,400 cal yr BP, material was deposited at the site by fluvial processes. The base of the core, from 312 cm to 280 cm, is made up of sand and organic material. The transition from sandy fluvial deposits to peat formation coincides with the Younger Dryas. For the purposes of discussion, the age-depth chronology of the Whiskey Fen core will be assumed to be linear.

Organic Matter and Bulk Density

Organic matter and bulk density (Figs 5.2, 5.3) have a generally inverse relationship with each other. Bulk density slightly decreases and organic matter content slightly increases from the beginning of the record in the Late Pleistocene and Younger Dryas, to the beginning of the Early Holocene 11,700 cal yr BP. The organic content values are less than 30%. Cooler conditions during the Younger Dryas likely limited the production of organic material at the site. Organic matter content increased during the Early Holocene, with a peak of about 78% organic matter at about 8,900 cal yr BP. The increased organic content is likely due to warming temperatures. From about 8,000 cal yr BP to about 6,500 cal yr BP (120-100 cm depth), organic matter content continues to increase to over 80% while bulk density also slightly increases. Organic matter content sharply declines while bulk density slightly increases after about 6,500 cal yr BP, with organic matter values below 20% for the first time since the Younger Dryas. While organic matter content can act as a proxy for temperature, drier conditions can negatively impact the production and preservation of peat. Temperatures continued to warm from the Early Holocene through the Middle Holocene, but conditions may have dried to the point of disrupting organic matter content as a temperature proxy during the second half of the Middle Holocene. As Middle Holocene temperatures increased, snowmelt occurred earlier, allowing the late summer water table to lower to the point that peat was no longer preserved, and fluvial processes dominated at the site.

Organic matter content increased back to peaks as high as 80% by about 4,700 cal yr BP, and remains relatively high throughout the remainder of the record.

Humification

High values for transmissivity generally indicate that peat is well preserved, and corresponds with periods of high water table. The high transmissivity values indicate that conditions were cooler and wetter. Low transmissivity values usually indicate poor peat preservation, and periods of lower water table, brought on by warmer and/or drier conditions. High transmissivity values, in excess of 100% corrected transmittance, may occur when peat is not preserved, so that there is insufficient organic matter in the sample from which to extract humic acids. Corrections for organic matter content to transmissivity values correct samples that do not have enough organic content to qualify as peat to values above 100%. Generally, higher transmissivity values indicate greater peat preservation, and greater decomposition at lower values.

Organic content values are all less than 40% from 190 cm to about 165 cm (10,600 cal yr BP). Peat formation was not yet occurring during the Younger Dryas and the beginning of the Early Holocene. High transmissivity values between about 165 and 175 cm depth occur at a time when organic matter values are low, indicating that there is little humic decomposition to affect transmissivity.

The discussion below refers to the humification data from 10,000 yrs ago to the present.

Between about 10,000 and 9,000 yrs ago transmissivity values are high. This period corresponds with cool and moist early Holocene conditions. Beginning around 9,000 yrs ago and extending until 7,000 yrs ago transmissivity values are generally lower, except for a short peak around 8,200 yrs ago. Climate reconstructions based on changing orbital parameters suggest that summer temperatures in the Northern Hemisphere were warmer from about 9,000 to 7,000 yrs ago (Kutzbach and Guetter, 1986). The low transmissivity values in this section of the Whiskey Fen core most likely correspond with this period of warmer summers, which would have resulted in earlier snowmelt and lower late summer water tables. The exception to this warming appears to be the spike in transmissivity values which probably corresponds to the 8,200 year event, a decades-long cooler period brought on by the catastrophic draining of Lake Agassiz-Ojibway (Barber *et al.*, *1999*; Cronin *et al.*, 2007*)*.

From about 8,200 yr BP until about 6,500 yr BP was a period of decreasing peat preservation. The increasing organic matter between about 165 and 130 cm depth probably correspond with warmer conditions. As the temperatures increased, snowmelt would have occurred earlier in the summers, and would have led to lower water tables toward the end of the summer, leading to less well-preserved peat.

From about 70–100 cm depth in the core organic matter reaches its lowest values of the Holocene, while bulk density values reach their highest. This zone also shows high transmissivity values. High transmissivity values between about 70 and 100 cm depth correspond to the period of low organic content and higher bulk density during the Middle to Late Holocene. Previous research on lower elevation fens in Colorado (< 3,100 m elevation) shows that during the Middle Holocene, warm and dry summer conditions prevented peat from accumulating in topogenous fens (Vierling, 1998; Sullivan, pers. comm.). Instead, fluvial sediments composed of interbedded silts, sands and organic mud accumulated in these areas.

Lower transmissivity values in the upper 75 cm correspond to the higher organic content as peat is re-established in the upper part of the core, from about 2,000 yrs ago to present. A period of low transmissivity values from about 30–42 cm would seem to correspond with the Medieval Climate Anomaly, a period of slightly elevated temperatures lasting from about 950 to 1250 CE. The higher temperatures would have resulted in lower water tables and more humified peat. This period was followed by higher transmissivity values between about 10 and 20 cm depth, which probably represent the Little Ice Age, a period of cooler temperatures lasting from about 1300-1850 CE.

Carbon Storage

Organic carbon storage (kg/m^2) increases with deeper peat depth intervals. Despite decreased organic carbon per square meter, the majority of organic carbon stored within the fens is stored in areas with less than 50 cm of peat depth, even when conservatively estimating an average depth of 25 cm for these areas, due to the large area of shallow peat in each of the fens. Pancake Fen contains an estimated total of 179.08 tons of organic carbon, Hatchet Fen contains an estimated total of 3,686.77 tons of organic carbon, and Whiskey Fen contains a total of 10,724.05 tons of organic carbon.

Carbon storage is much greater in the peat of Whiskey Fen than the geographically-estimated or wetland type-estimated values produced in Fennessy and Nahlik from the 2011 National Wetland Assessment (2016; Table 6.1). The top 120 cm of Whiskey Fen contained 163.483 kg/m^2 of organic carbon, while Fennessy and Nahlik's estimated carbon storage from the 2011 National Wetland Assessment ranges is 18.59–27.77 kg/m^2. A conservative estimate of the volume of peat within the three surveyed fens indicates that Pancake Fen has the smallest amount of carbon storage at 179.08 tons, followed by Hatchet Fen (3,686.77 tons) and Whiskey Fen (10,724.05 tons) (table 5.3). As a frame of reference, the typical passenger vehicle emits about 4.7 tons of carbon dioxide per year

(EPA, 2016). Whiskey Fen conservatively stores the equivalent of about 2,282 passenger vehicles' emissions over one year. The disparity between the estimated organic carbon storage and the conservatively calculated organic carbon storage from the Whiskey Fen core underscores the need for additional study of mid-latitude peatlands.

The largest amount of carbon storage in the three fens occurs within peat less than 50 cm in depth, despite the highest carbon density occurring in deeper sections. This is due to the majority of the area of fens being relatively shallow in peat. This is significant, because the upper portion of peat is within or near the acrotelm and susceptible to decomposition during periods of decreased moisture (Whittington and Price, 2006). A small decrease in saturation level can initiate a positive feedback of diminishing carbon storage ability by allowing for compaction. Compaction leads to decreased porosity and hydraulic conductivity of the peat. The compromised functioning of subalpine fens could have many negative impacts on water quality and supply. Soil carbon storage modeling conducted by Ise *et al*., (2008), and summarized in section 2.3, found an 86 % loss in carbon storage within a peatland experiencing a 4° C temperature increase. Subalpine peatlands help regulate the flow of snowmelt from high elevations, and decrease the sediment transport of overland flow.

The two largest gaps in peat preservation at Whiskey Fen occur at the base of the core (161–190 cm, 12,400–10,000 cal yr BP) and from 66–101 cm (6,000–1,600 cal yr BP). Both of these intervals correspond to periods of warming shown in other paleoenvironmental records in the region, and the latter period with a period of warming and drying in the Harbison Pond record as discussed below. The timing of these gaps in peat formation suggests that the carbon sink function of the subalpine fens in the Colorado Rockies may be particularly sensitive to future warming.

Table 6.1. Calculated organic carbon storage per square meter in Whiskey Fen compared to estimated organic carbon storage of wetlands in the West and estimated organic carbon storage for emergent wetlands using the 2011 National Wetland Condition Assessment (Fennessy and Nahlik, 2016).

Whiskey Fen	**Organic Carbon Storage (kg/m^2)**	**Fennessy and Nahlik, 2016**	**Organic Carbon Storage (kg/m^2)**	**Fennessy and Nahlik, 2016**	**Organic Carbon Storage (kg/m^2)**
Calculated 0–10 cm	6.62	West Estimate 0–10 cm	1.92 ± 0.22	PRL-EM Estimate 0–10 cm	3.47 ± 0.22
Calculated 0–30 cm	26.22	West Estimate 0–30 cm	5.67 ± 0.63	PRL-EM Estimate 0–30 cm	1.01 ± 0.71
Calculated 31–60 cm	36.47	West Estimate 31–60 cm	5.32 ± 0.68	PRL-EM Estimate 31–60 cm	8.26 ± 1.04
Calculated 61–90 cm	41.56	West Estimate 61–90 cm	5.34 ± 0.82	PRL-EM Estimate 61–90 cm	5.49 ± 1.23
Calculated 91–120 cm	59.24	West Estimate 91–120 cm	5.3 ± 0.91	PRL-EM Estimate 91–120 cm	4.68 ± 1.66
Total 120 cm (kg/m^2)	**163.48**	**Total (kg/m^2) West Estimate for 120 cm wetland**	**21.63 ± 3.04**	**Total (kg/m^2) PRL-EM Estimate for 120 cm wetland**	**22.91 ± 4.86**

Harbison Pond

Core Description and Chronology

The oldest retrieved radiocarbon date from the Harbison Pond core is 16,201 cal yr BP at 350 cm. The maximum extent of the Pinedale glaciation in the Kawuneeche Valley and Middle Park has been thought to have been the moraines located at the southern end of present-day Shadow Mountain Reservoir, now visible as narrow islands (Richmond, 1974; Figure 3.3; Figure 6.1). The Last Glacial Maximum of the Pinedale glaciation occurred about 19,000 yr BP in western North America (Richmond, 1974; Owen *et al*.,2003; Benson *et al*.,2005; Clark *et al*., 2009). Dating of glacial features elsewhere in and near the Colorado Rockies indicate that deglaciation after the Pinedale proceeded relatively rapidly after about 17,000 yr BP to 14,000 yr BP, with a small glacial advance occurring during the Younger Dryas about 10,000 yr BP (Reasoner and Jodry, 2000). The Colorado River Glacier occupied the broad and south-facing Kawuneeche Valley, which adds additional plausibility to a relatively early and rapid retreat.

The basal date of the Harbison Pond core is very likely influenced by contamination, given the consistent older chronology of samples from elsewhere in the core. Using the radiocarbon date from 350 cm, Harbison Pond began collecting sediment in the late Pleistocene, at least more than 16,201 cal yr BP. If the moraines at the southern end of Shadow Mountain Reservoir were deposited during the Pinedale Glacial Maximum, the Colorado River Glacier would have been retreating an average of about 4.1 m a year from about 17,000 to 16,201 yr BP. The age of the kettle pond suggests that rather than the Pinedale Glacial Maximum forming the Shadow Mountain Reservoir moraines, they may have been deposited by the Bull Lakes glaciation more than 130,000 years ago and that the Pinedale glaciation was not as extensive in the Grand Lakes region as previously thought. Additional mapping and dating of glacial features in the area is necessary to better understand how Harbison Pond and the morraine on which it formed fit into the glacial history of the region.

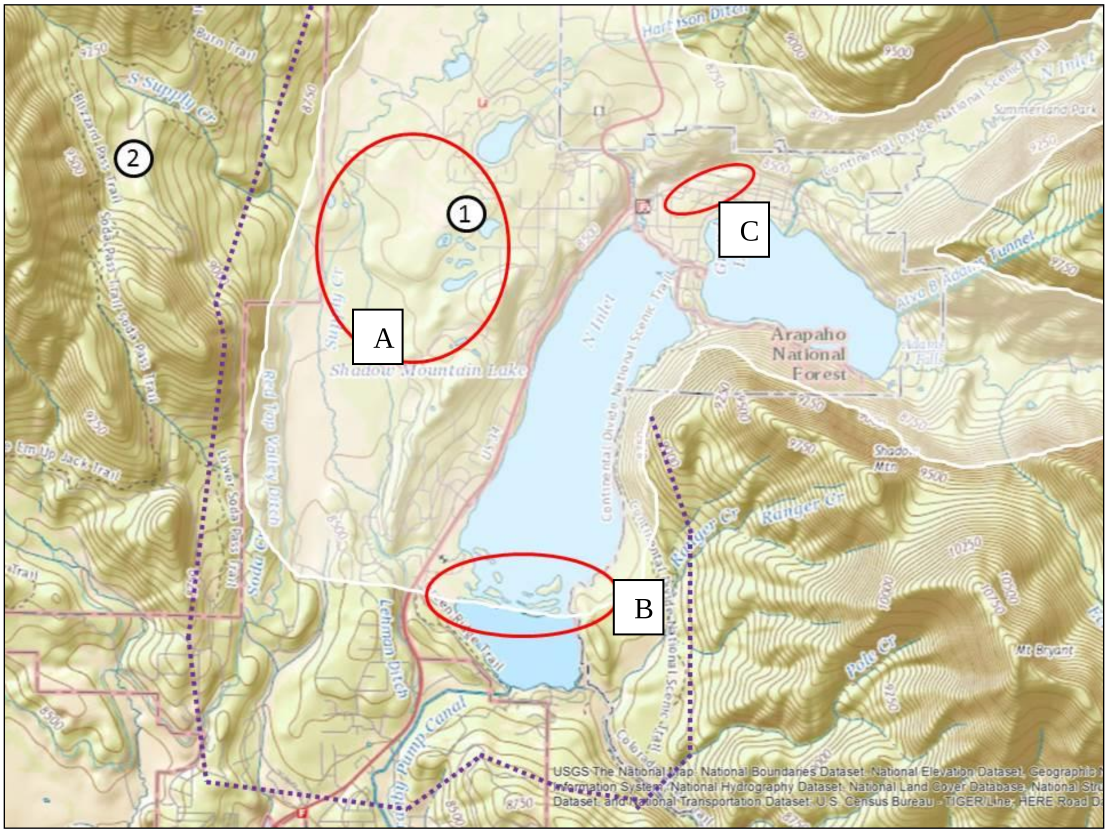

Figure 6.1. Glacial features and Richmond's (1974) glacial extent around Grand Lake, CO (see also Figure 3.1). The White line and faded area is the estimate of the extent of the Pinedale glaciation given by Richmond (1974), and the dark dotted line is the estimate of the extent of the Bull Lake glaciation. 1: The location of the sediment coring site, Harbison Pond. A: Hummocky, morainal terrain with several undrained depressions and ponds. B: Terminal or recessional moraines that have become islands by the creation of Shadow Mountain Reservoir. C: A prominent lateral moraine within the city of Grand Lake.

Magnetic Susceptibility and Organic Content

The low amount of organic content and high magnetic susceptibility of the sediment from the bottom of the record until about 7,250 cal yr BP in the first half of the Middle Holocene indicates that relatively little autochthonous material was being deposited during this time. The laminations of fine glacial flour and sand suggest periods in which a larger portion of sediment influx was windblown (glacial flour) and other periods in which the influx was overland flow during heavy rain or rapid snowmelt. A regular influx of fine glacial flour may have decreased water clarity and further hindered production within the lake of organic sediments.

During the second half of the Middle Holocene, the magnetic susceptibility of sediments begin to decline and organic matter content begins to increase. At the beginning of the Early Holocene, magnetic susceptibility sharply decreases and organic matter content increases. The increasing autochthonous input suggests that conditions remained relatively warm. This shift in productivity within the lake may be due to lower lake levels allowing for better light penetration and better habitat for aquatic plant species.

Macroscopic and Microscopic Charcoal

The small size of Harbison Pond's catchment ensures that the macroscopic charcoal signal represents an area of only a few hundred meters around the pond. Macroscopic charcoal in the sediment could not have originated more than a few hundred meters away from the edge of Harbison Pond at present, as macroscopic material cannot travel through the air long distances and is usually deposited by being washed in by overland flow of water. Macroscopic charcoal has been shown to disperse through the air to a limited degree, up to 2 km from a fire event (Tinner *et al*., 1998). Overland flow washes charcoal either formed or deposited within a basin into the sediment of a catchment, making the extent of a lake's basin a major influence on the sediment record (Whitlock and Millspaugh, 1996). While the geographic scope of Harbison Pond's macroscopic charcoal record is narrow, the lack of potential transport of distant charcoal from overland flow minimizes the potential local bias of much of the microscopic charcoal record. This is likely why the FRI is 1,000 to 3,000 years, while studies within Rocky Mountain National Park with a larger catchment area have calculated Holocene FRI ranging from 150 to 500 yrs (Veblen *et al*., 1994; Buechling and Baker, 2004; Caffrey and Doerner, 2012).

The decrease in charcoal fragments within the record during the Middle Holocene may also be due to less fuel availability in the area, or simply be relative to extremely high fire activity in the early Holocene. Another possibility is that the immediate area surrounding Harbison Pond was not forested during the Middle Holocene, reducing potential fuel and charcoal sources for this small basin. Fall (1997) found that lower treeline retreated upslope as much as 200 m in the Colorado Rockies during the middle Holocene, due to drier conditions. Harbison Pond is presently near the ecotone between the montane forest and sage steppe in north Middle Park. Well-drained, south-facing slopes at similar elevations of Harbison Pond are unforested 7 km south of the study site, and are thinly wooded less than 2 km south of the study site. A lower treeline retreat of 100–200 m wood likely remove the lodgepole pine stand presently occupying the lateral moraine on which Harbison Pond is located.

Periods of increased fire activity could correspond to periods of low moisture availability in the Whiskey Fen record, and there are some relationships between the humification record from Whiskey Fen and the charcoal records from Harbison Pond. The period of lowest transmissivity, and thus a period of variation in the saturation level of Whiskey Fen, is centered at 7,500 cal yr BP. This corresponds with the highest peak in microscopic charcoal from Harbison Pond, also at about 7,500 cal yr BP. It also falls within the period of the macroscopic charcoal record with the highest peaks and calculated fire frequency. However, the sampled levels that are above 100% transmissivity may represent the actual periods of lowest moisture availability in the fen record, in that the material deposited at that time was left exposed to air and decomposed. The first interval of Whiskey Fen sample levels without peat formation, 161–190 cm (11,200–12,347 cal yr BP) does not date within the recoverable pollen record. The second interval of excluded levels, 66–101 cm (4,500–6,400 cal yr BP), occurs during a period of declining fire activity immediately following the period of time with the most frequent and intense fire activity. Above, I suggested this decline was due to a decrease in available fuel following the onset of drier conditions after a period of relatively frequent and intense fires. This interval of low soil saturation supports the idea that the middle Holocene was much drier than present, and may have inhibited fuel development for fires.

Both the microscopic and macroscopic charcoal records indicate increased fire activity in the Early Holocene and first half of the Middle Holocene, from 8,500–6,900 cal yr BP. Somewhat drier than present summer conditions, or more interannual variability in precipitation may have allowed for both the high fire frequency and intensity that is indicated by the charcoal record. Fire events and sparse vegetation on young soils may have combined to allow for increased wind-born transport of fine material like that which dominates the bottom 162 cm of the core. The three large peaks in microscopic charcoal correspond with peaks within the macroscopic record, and are likely influenced by fires occurring in the immediate vicinity of Harbison Pond.

Low macroscopic and microscopic charcoal in the first half of the Late Holocene, the lowest in the record, indicates that there was less fire activity during this time than present or at any other time during the Holocene.

Surface Pollen

The elevational transect of modern surface pollen reflects the relative complacency of the Harbison Pond record, but does provide some insights as to which taxa are most responsive to geographic shifts of plant communities. *Sarcobatus* low-spine Asteraceae, and Poaceae taxa are more numerous at lower elevations and are likely more local signals when present with the Harbison Pond record. *Picea* taxa was only abundant above 2,700 m. *Pinus* taxa peak above 2,600 m elev, but are overwhelming dominant at all sampled elevations, including far below treeline. High-spine Asteraceae presence appears to be more site-specific than all other taxa, and not correlated to elevation. This is likely due to the wide variety of plants within the family Asteraceae with similar pollen grain morphology, and because Asteraceae are zoophilous.

Pollen Analysis

The pollen record from Harbison Pond is somewhat complacent (Figures 5.16, 5.17, 5.18), with *Pinus* pollen likely overwhelming signals from less abundant taxa. *Pinus* grains are able to disperse long distances, can be present in multiple vegetation zones, and are produced in large quantities (Fall, 1992; Fall, 1997). However, combining the pollen record with other proxies provides additional context that aids in interpreting changes in the record over time.

The Harbison Pond pollen diagrams are divided into four pollen zones that reflect changes in significant pollen taxa. Pollen preservation within the core was adequate for identification and counting only on levels taken from the upper portion of the core from 0–245 cm, with gaps in countable levels between 195 cm and 245 cm. The presence of glacial flour in sediment below 155 cm obscured grains when examining them through the microscope. Preparing additional slides and increasing the number and duration of hydrofluoric acid (HF) treatments during pollen allowed for levels with only moderate amounts of glacial flour below 155 cm to be counted.

Pinus pollen remain dominant throughout the record. Attempts to differentiate between haploxylon and diploxylon were made, but the presence of glacial flour made many grains indeterminable that otherwise likely would have been identifiable. Haploxylon grains have a membrane with small papillae, while diploxylon grains have a smooth membrane. Identifying whether a *Pinus* grain was haploxylon or diploxylon can be difficult due to the need to roll the grains into a polar view. Haploxylon *Pinu*s grains, most like *Pinus* edulis, were identified in samples from earlier in the record, but are absent from samples near the top of the core. A large majority of the indeterminate *Pinus* grains were likely diploxylon grains, but I was unable to maneuver the grains to confirm this.

Pollen zone IV (195-245 cm)

This pollen zone includes the deepest samples in the core, and includes samples dating to approximately 15,000 to 15,800 years ago. *Artemisia* and *Picea* are abundant in this section of the record. Cooler and drier conditions during the late Pleistocene would have lowered treeline in the region, leaving the Harbison Pond site somewhat above or near treeline. High values of *Artemisia*, Poaceae and Asteraceae reflect the proximity of alpine tundra vegetation. Spruce trees were probably part of the upper treeline forest. The relatively low pine pollen values indicate that pine stands were depressed to lower elevations, not near the core site.

Pollen zone III (152-195 cm)

This zone includes the late Pleistocene and early Holocene. Sedimentation rates were low in this period. Pine pollen becomes dominant in this zone. In lower Zone III diploxylon pines, presumably either *P. ponderosa*, or *P. contorta*, become established in the area. *Picea* percentages drop from those in Zone IV, but spruce is probably present in the higher elevation sites near the lake. *Artemisia* values increase in this zone, and may represent the spread of *A. tridentata* as the climate warms and dries. Increases in

Sarcobatus, Poaceae, and Amaranthaceae pollen in this zone would tend to suggest warming and drying, as well.

Nuphar pollen increases during zone III, perhaps signaling a shallowing of the lake in response to the warmer and drier conditions.

Other paleoenvironmental studies have found that the early Holocene was as warm or warmer than present, with treeline rising from Pleistocene elevations (Elias, 1996; Fall, 1997). The relative abundance of *Picea* taxa may be due to this upslope expansion of the subalpine forest, with a slower succession of subalpine to montane forest at lower elevations as communities shifted in response to climate. Warmer temperatures would likely mean more frequent drought events and drier conditions at lower elevations, which is consistent with the prevalence of *Artemisia* within the early portion of the record. This section of the core has the lowest pollen concentration, which could be due to less pollen production overall due to stress from warmer and drier prevailing conditions to which plant communities had not yet fully shifted in response.

Pinus percentages somewhat increases over the section of the core. *Pinus* is relatively complacent for most of the record, likely due to the transport range and prodigious pollen production of *Pinus* taxa compared to other plants. This creates a strong regional signal less impacted by natural climate variability on a millennial scale. However, the increase in abundance of *Pinus* taxa from the Early Holocene into the first half of the Middle Holocene may be due to the expansion of the montane forest in response to warmer and wetter conditions.

Pollen zone II (50-152 cm)

Pinus and other arboreal pollen reach their highest values in this zone. Haploxylon pine, most likely *P. edulis*. becomes more common in this zone. On the other hand, percentages for *Artemisia, Sarcobatus*, and high spine Asteraceae fall to their lowest Holocene values, while Poaceae and Amaranthaceae increase. This zone represents mid-Holocene conditions when many researchers have noted a drying trend. The pollen data suggest that an open pine woodland occupied the site.

Nymphaceae (*Nuphar polysepala*) pollen is present throughout the record, but greatly increases in Zone II. The increase in Nymphaceae pollen coincides with changes in magnetic susceptibility and organic content in the second half of the Middle Holocene. *Nuphar polysepala* has a preference for water depths between 1.5 m and 2 m. The flat-bottom bathymetry and relatively steep edges of Harbison Pond would make the area of the lake with that level of depth much smaller than present with any large fluctuation in lake levels. The increase in Nymphaceae abundance in this zone is further evidence of a decrease in lake depth during this time that culminated during Zone II.

Other paleoenvironmental studies in the region have also found evidence of drier conditions during this time. A decline in subalpine pollen taxa at sites in Wyoming and Idaho during the Middle Holocene indicate drier and warmer conditions in the Rockies north of Harbison Pond. (Doerner and Carrara, 2003; Whitlock, 1993; Whitlock *et al*., 1995). Speleothem records from the Wasatch Mountains in Idaho indicate winter precipitation declined during the middle Holocene, which would lower moisture availability during the late spring and summer and exacerbate drought events (Lundeen *et al.*, 2013).

Pollen zone I (0-50 cm)

The increasing organic matter content of the lake sediment suggests that conditions remained relatively warm during the Late Holocene, but Nymphaceae taxa declines somewhat in this section of the core. Cyperaceae and *Typha* taxa, while still absent from many sample levels, reach their highest abundances within the record during this time period. At present, little Cyperaceae or *Typha* is within or on the shore lake. However, Cyperaceae and *Typha* are abundant within and around many of the neighboring kettle lakes. An increase in water level in the relatively flat-bottomed pond might reduce the area within the preferred water depth range of *Nuphar*, as well as create shallow areas (less than .5 m depth) in the northeast and southwest corners that would be suitable for Cyperaceae and *Typha*. *Artemisia* taxa declines during this period while *Pinus* pollen, remains dominant in the record. The relative decline of *Artemisia* and high values for *Pinus*, Cyperaceae, and *Typha may* indicate wetter conditions in the Late Holocene.

Comparison with other Studies

The past conditions indicated by the Whiskey Fen and Harbison Pond records largely agree with the general climate trends identified other paleoenvironmental studies in the region (Table 6.2). The southern Rockies were cooler and drier than present during the Late Pleistocene. The Colorado Rockies during the Early Holocene were generally warm and wet, though there is some variation in conditions within the broader region. Conditions in the region during the Middle Holocene were generally warmer and drier, and generally cooler and wetter in the Late Holocene.

Table 6.2. Summary of general climate trends in studies conducted in the southern Rocky Mountain region.

Study and Location	**Late Holocene**	**Middle Holocene**	**Early Holocene**	**Late Pleistocene**
Benedict *et al*., 2008; Colorado Rockies	Cool	Warm	-	-
	-	-	-	-
Bright and Davis, 1982; Snake River Plain, Idaho	Cool	Warm	-	-
	Wet	Dry	-	-
Caffrey and Doerner, 2012; Colorado Rockies	-	Warm	-	-
	-	-	-	-
Carrara and McGeehin, 2015; Colorado Rockies	Cool	-	-	-
	-	-	-	-
Doerner and Carrara, 2001; west-central Idaho	Cool	Warm	Warm	Cool
	Wet	Dry	Dry	Dry
Elias *et al*., 1996; Colorado Rockies	Cool	Cool	Warm	Cool
	-	-	-	-
Fall, 1997; Colorado Rockies	Cool	Warm	Warm	Cool
	-	Wet	Wet	Dry
Lundeen *et al*., 2013; Southeast Idaho	Cool	Warm	Cool	-
	Wet	Dry	Wet	-
Reasoner and Jodry, 2000; Colorado Rockies	-	-	Warm	Cool
	-	-	-	-
Vierling, 1998; Colorado Rockies	Cool	Warm	Warm	Cool
	Dry	Dry	Wet	Dry
Whitlock *et al*., 1995	Cool	Warm	Warm	Cool
	Wet	Dry	Dry	Dry
Harbison Pond and Whiskey Fen Records	**Cool**	**Warm**	**Warm**	**Cool**
	Wet	**Dry**	**Wet**	**-**

7. SUMMARY AND CONCLUSION

The objectives of my research were to: 1) reconstruct vegetation and climatic changes through the Holocene, 2) reconstruct changes in sedimentation patterns, sedimentation sources, and *in situ* hydrology and 3) use the information from the first two objectives to assess the scope of possible future impacts from anthropogenic climate change to wetland function, water quality, and water availability in western rivers. Based on the three objectives above, I formulated six research questions. These research questions and brief responses are below:

How stable are modern vegetation communities in the central Southern Rockies?

Although the Harbison Pond pollen record is somewhat complacent, the pollen data indicate that modern vegetation communities are the products of dynamic responses to climate variation and disturbance, and are continuing to respond to these long-term patterns in the present. A change in fire activity as observed through the microscopic and macroscopic charcoal records also occurred during the Holocene, as fire frequency greatly declines due to warmer, drier conditions limiting fuel availability. The pollen record is generally complacent with *Pinus* taxa overwhelming other signals for much of the record.

To what extent and in what way is evidence from past climatic events such as the Younger Dryas, Holocene Climatic Optimum, and the Little Ice Age preserved in the record?

The pollen record from Harbison Pond is not as sensitive as anticipated, and provides relatively little evidence for these climatic shifts, although changes in lake levels are seen in the aquatic pollen record. There is a decline in nonarboreal pollen taxa that coincides with the Little Ice Age, but no other noticeable response in the records.

How have subalpine fens responded to variations in global temperature in the past as indicated by oxygen isotope records?

Changes in the biogeochemical properties of the Whiskey Fen core provide a very sensitive record of past hydrologic change. The humification record clearly shows changes in effective moisture consistent with the cooler conditions of the 8200-year event and the Little Ice Age, while the warmer conditions during the early Holocene Climatic Optimum and the Medieval Climate Anomaly are registered in the poorer peat preservation and low transmissivity of peat accumulated during these periods. Furthermore, a hiatus in peat formation from about 2,000 to 6,500 years ago corroborates similar findings from other lower elevation subalpine fens in central Colorado. This finding is quite significant as it indicates that these wetlands are particularly vulnerable to relatively small increases in summer temperatures. Considering the importance of these fens in carbon sequestration, water quality and streamflow, and biological species diversity, this is a matter of considerable concern for the future.

How might changes in mid-latitude subalpine fens affect water resources, as well as carbon storage and sequestration?

Whiskey Fen responded to warmer and drier conditions in the past by decreased organic carbon storage and peat preservation, signaling a decrease in water table level. Lower soil saturation levels lead to compaction, which diminishes the porosity and hydraulic conductivity of underlying material and forms a positive feedback loop towards further decreases in soil saturation levels. The role subalpine fens play in water storage and flow regulation, as well as associated impacts on erosion control, would be diminished or inhibited. Unsaturated peat decomposes and releases carbon, lowering carbon storage. The decreased porosity and hydraulic conductivity hinders the continued development of peat, decreasing carbon sequestration.

Are there any analogs for future climates in the record that may be used to predict future vegetation change in the area?

The Holocene Climatic Optimum was warmer in the Colorado Rockies than present conditions, and can serve as an analog for the warming conditions expected by 2050. In addition, the drier conditions from 2,000–6,500 years ago had a profound effect on peat accumulation and preservation, and indicates that future warming and drying in the Colorado Rocky Mountains may result in a substantial decrease in the area of peatlands in the lower subalpine and montane zone, with consequent declines in carbon storage and water quality.

To what extent are human impacts on vegetation, such as grazing, fire, or agriculture, detectable in the record?

Amaranthaceae and Poaceae taxa are presently higher relative to the recent past, but there are analogous fluctuations throughout the record. There is no indication of human influence on fire within the record.

The timing of glacial retreat in the Colorado Rockies at the end of the Pleistocene was extremely rapid. The colonization and development of modern vegetation communities began simultaneously with this retreat, but may have taken centuries to reach full cover. The broad range of elevations and environments in which *Pinus* taxa can successfully survive, coupled with prodigious pollen production and long travelling distances, make it the dominant contributor to the pollen record through several climatic events and presents a challenge to reconstructing past environments within the Colorado Rockies. Combining the pollen record with additional proxies allows for small variations in other taxa to be seen within a larger context and contribute to interpreting past conditions.

Fire has always been a major contributor to the landscape of the Kawuneeche Valley and the Colorado Rockies. The Early Holocene experienced more frequent and intense fires than present. As climate continued to warm and dry during the Middle Holocene, the growth of fuel slowed and fire became somewhat less frequent around Harbison Pond. After the Holocene Climatic Optimum and Middle Holocene, fire was again a major part of the ecosystem, but remains diminished in severity and frequency compared to the Early Holocene.

Harbison Pond underwent a major change during the Middle Holocene. Warm temperatures and dry conditions led to a drop in lake levels that transformed Harbison Pond from an open surface lake receiving mostly wind-borne and water-borne material to a vegetated lake accumulating mostly autochthonous organic material. These changes coincide with a pause in peat formation in Whiskey Fen. Other research has shown that warmer conditions can lead to positive feedback loops that diminish or destroy the ability of midlatitude peatlands to accumulate and preserve peat. Mid-latitude subalpine fens require more research attention in the face of anthropogenic climate change in the West.

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APPENDICES

Appendix A. Pollen Processing Procedure

After Berglund and Ralska-Jasiewiczowa, 1986.

1. Wash and rinse with deionized water six numbered Nalgene test tubes, and place them in a color-coded test tube rack.

2. Wash and rinse the 1 cm^3 ceramic sampling spoon and the small spatulas with deionized water.

3. With the back of the spatula, carefully scrape the most superficial layer of sediment at the face of the split core to one side in one motion. With the other end of the spatula, dig out approximately 1 cm^3 of sediment and press it into the sampling spoon compactly, smoothing the top level to be even with the top of the spoon.

4. Place 1 cm^3 of sediment into each test tube, recording sample depth in each test tube and the color of the rack.

5. Add 1 *Lycopodium* tablet to each sample.

6. Partially fill each test tube with 10% hydrochloric (HCl), and allow the liquid to react with the tablet. When the bubbling subsides, fill the test tube with more HCL and stir the sample with a wooden applicator gently. When the reaction has subsided, stir the sample a little more vigorously and try to loosen up the sediment. Allow the samples to sit in HCl under a fume hood overnight. Before proceeding, the sample should no longer be reacting.

7. Begin heating a hot water bath for Step 8. Centrifuge the samples for 2 minutes at about 2800 rpm, then decant the supernatant into the plastic reagent container under the fumehood. Fill the tubes with deionized water, stir, centrifuge, and decant, and then again fill the tubes with deionized water and repeat.

8. Begin heating a large beaker of hot deionized water for Step 9. Add 5% potassium hydroxide (KOH) to deflocculate clays and disperse organics. Once the hot water bath is gently rolling, place the test tubes in a rack submerged in the bath, stirring two or three times over 8 minutes.

9. Centrifuge the test tubes and then decant the reagent into a container. Rinse the sample with hot deionized water and then centrifuge. Repeat these steps until the supernatant is clear. If the samples are not becoming clear after more than 12 rinses, proceed to the next step and then return to centrifuging, decanting, and rinsing until it becomes clear. With samples containing glacial flour, it can take more than 20 rinses.

10. Place six 125µm sieves in an ultrasonic cleaner for 20 minutes, and then rinse them with deionized water. Nest the sieves over numbered plastic beakers, placed in numerical order. Swirl and then pour each test tube on to a sieve, then lightly spray a small amount of deionized water from a water bottle across the samples. Once the samples are thoroughly rinsed, set aside the sieves and transfer the sample that has collected in the beakers back into the same test tubes. Centrifuge and decant. If the supernatant is not clear, continue centrifuging, decanting, and rinsing until it is.

11. Put on an HF-certified respirator, protective apron, and face mask. Put on latex gloves, and then rubber gloves over them. HF stands for "hydrofluoric acid," which should never be touched or left out. Neutralize any spills with baking soda, which should be kept very handy.

12. Carry the samples to the HF hood. Carefully fill the test tubes half way with HF with the tubes pointed away from you, and slowly stir. Allow the samples to sit overnight, but no more than 48 hours.

13. Upon returning, be sure to wear all the safety gear again. Centrifuge and decant the supernatant into a plastic HF waste container containing plenty of baking soda. The supernatant will react, and this is normal.

14. Rinse the samples with 10% HCl then stir, centrifuge, and decant. Now rinse the samples with deionized water twice, stirring, centrifuging, and decanting each time.

15. Carefully check that there are no spills on or under the test tube rack. If there are, carefully neutralize with baking soda while keeping your test tubes covered in plastic wrap. When safe, transfer back to the other fume hood. Be sure to clean up all traces of your HF step before leaving, as a matter of safety.

16. Rinse the samples once with hot water, and check a small subsample under a microscope. If there is a great deal of organic material, allow the samples to sit in 5% nitric acid (HNO_3) for 5 minutes and then centrifuge, decant, and rinse twice with deionized water.

17.Prepare a 9:1 mixture of acetic anhydride and sulphuric acid (H_2SO_4) by adding 6 ml of sulphuric acid to a graduated cylinder with 54 ml of acetic anhydride. Note that this mixing should be done in the order stated and not the other way around. Prepare a hot water bath. Pour 8 ml of glacial acetic acid into each test tube, stir, then centrifuge and decant. Note: DO NOT RINSE.

18. Carefully pour the acetic anhydride and sulphuric acid mixture into each test tube and gently stir. If the samples his and spit, you forgot to rinse with glacial acetic acid or otherwise let water be present. Note that the mixture will react to the stick. This is normal. Place the samples in the hot water bath for about 9 minutes, stirring once or twice.

19. Centrifuge, decant, and rinse with glacial acetic acid. Centrifuge and decant, then rinse the samples with hot deionized water. After again centrifuging and decanting, fill the tube half way with 5% KOH and then top up with hot water. Place the samples in a hot water batch for 5 minutes.

20. Centrifuge, decant, and rinse with hot deionized water at least two times. The supernatant should be clear.

21. Prepare a safranin stain solution by adding 2-3 drops of 1% safranin solution to 100 ml of deionized water. Fill each test tube with the solution, stir, centrifuge, and decant. Rinse once with water.

22. Transfer the samples to rinsed, etched vials with labeled corks. Do this by adding about 1 ml of tertiary butyl alcohol (TBA) into each tube, stirring, and then pouring the samples into the appropriate vials. Be sure not to come in contact with the TBA. Repeat if necessary, centrifuging and decanting the vials if they become overfilled. When finished, centrifuge and decant the vials.

23. Add about two drops of silicone oil (2000 centistokes viscosity) to each vial and stir vigorously (but carefully). Stir each sample again every 30 minutes for the next hour. Leave the vials uncorked at least 24 hours so any remaining TBA will evaporate.

Appendix B. Macroscopic Charcoal Protocol

After Whitlock and Millspaugh, 1996.

1. Sampling should occur in numerical order (that is, work your way down a core, starting at
0-1 cm) and should be contiguous through the core.

2. Use an open syringe to take 3 cc samples from the core or bags of sediment. Make sure
the mud is packed tightly into the syringe so that air pockets are avoided.

3. Make sure a small beaker is labeled with the depth interval for the sample.

4. Empty the measured sample into the beaker, first by flipping the sample out of the syringe into
the beaker, and then by carefully rinsing any sediment that may still be stuck to the syringe with a squirt bottle filled with sodium hexametaphosphate. Make sure the syringe is completely clean.

5. Fill the small beaker with no more than 10 ml of 5% sodium hexametaphosphate (Na-HMP). Gently shake the beaker to disperse the sample.

6. Allow the sample to soak for 24–48 hours.

7. Add 5 ml of bleach to the beaker and gently shake. Soak the samples for 4 hours, periodically gently shaking and swirling the beakers. All of the sediment should unclump and take on a pale color, settling on the bottom.

8. Place a 250 µm sieve over a 125 µm sieve over a plastic beaker. Wash sediment through the top sieve as carefully as possible, working it down to a “pocket” at the bottom of the sieve. Using a
squirt bottle, carefully transfer the residue into a scored glass petri dish (larger lines). Place a lid
on the dish, make sure it is labeled with the same information as the sample beaker.

9. Wash the remaining sediment through the 125 µm sieve using the same technique as above.
Using a squirt bottle, carefully transfer the residue into a scored petri dish (smaller lines).
Place a lid on the dish, make sure it is labeled with the same information as the sample vial.
Stack this petri dish on top of the other petri dish and place it on a tray.

10. Place the glass petri dishes into a desiccator with their lids ajar and set the temperature to 90°C. Be sure to verify that the petri dishes are indeed glass and not plastic. Allow the petri dishes to dry for 4 to 6 hours.

11. Wearing oven mitts, carefully replace the lids of the petri dishes and remove them from the desiccator to cool for 30 minutes.

Making 5% Sodium Hexametaphosphate (Na-HMP):

1. Measure 50 g of sodium hexametaphosphate using the digital balance and a weighing tray.

2. Heat 1000 ml of dH20 on the hot plate.

3. Add sodium hexametaphosphate to the dH20, stirring until the crystals have dissolved.

4. Remove from hot plate and let cool.

5. Pour into appropriate squirt bottle.

Appendix C. Humification Procedure

After Chambers *et al*., 2011.

1. Sample 3 cm^3 peat (e.g. for very fibrous and/or very wet peat, use 1 × 1 × 0.5 cm; for less fibrous peat, use less) and place each sample in a pre-weighed crucible. Dry samples in crucibles in an oven (~95ºC) for 3 hours.

2. Grind up each peat sample separately in an agate pestle and mortar* and return to its crucible.
*Note: clean the pestle and mortar between each grind using dry paper or clean cloth—do not wash.

3. Weigh out 0.2 g of each sample accurately on a top-loading balance (to 3 decimal places) using a
watch glass and tare weight—then transfer to a 200-ml beaker (one beaker per sample). Process a batch of 12 samples at a time.

4. Turn on hotplate to preheat to 85–100°C.

5. Prepare 1.2 liter 8% NaOH solution. To do this, dissolve 96 g NaOH granules of AnalaR (or suprapur)
grade in about 600 ml deionized water and add additional deionized water up to 1.2 Liters.

6. Add 100 ml of 8% NaOH solution to each beaker using a 100-ml measuring cylinder.

7. Place beakers on the preheated hotplate and simmer at 95°C.

8. Top up beakers occasionally with deionized water to prevent drying out and to ensure solution does not become too concentrated.

9. After one hour of simmering, remove the beakers from heat and turn off the hotplate. Pour the contents of each beaker into a separate 200 ml labeled volumetric flask using a separate clean
funnel for each sample, and wash all residue into flask with deionized water.

10. Top up flasks to mark, and then stopper each flask and shake well.

11. Filter 50 ml of the contents of each 200-ml flask into the corresponding labeled 50 ml volumetric flask, using Whatman No. 1 grade filter papers.

12. Decant 50 ml of filtrate into labeled 100 ml volumetric flask and top up to 100 ml mark with deionized water. Stopper and shake well.

13. Turn on spectrophotometer to stabilize. Set to 540 nm. Calibrate with a sample cell filled to the marker line with deionized water.

14. When all flasks are ready, pipette a small volume from the first one into a sample cell and, using the spectrophotometer (previously allowed to stabilize), measure and record Absorbance and % light transmission of the first sample. Take three readings, and record the average. Between each reading, recalibrate the spectrophotometer with the sample cell filled with deionized water. Be sure to keep both sample cells wiped clean of fingerprints or other obstructions.

17. Repeat Step 14 for each sample.

Appendix D. Site Selection and Assessment

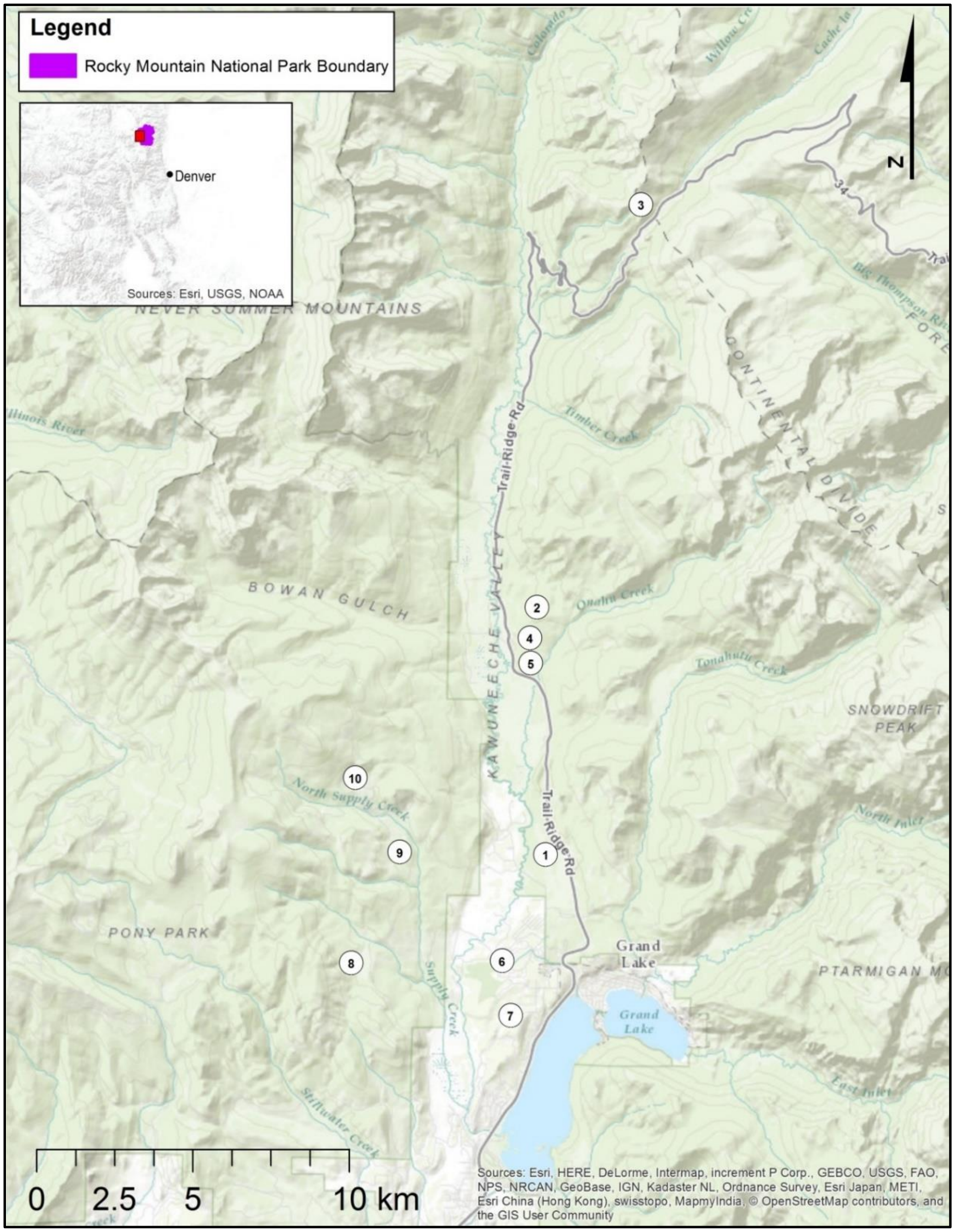

Figure D.1. Locations of surveyed sites. 1): Colorado River floodplain lake; 2) Chickaree Lake; 3) Bighorn Lake and nearby ponds; 4) Unnamed lake near Onahu Creek; 5) Onahu Creek fen adjacent to the unnamed lake 6) Columbine Lake 7) Six kettle ponds in Grand Lake Metropolitan Recreation District; 8) Whiskey Fen; 9) Pancake Fen; 10) Hatchet Fen.

Table D.1. Summary of surrounding vegetation and remarks from each survey location.

Location #	Name	Elevation	Surrounding Vegetation	Remarks	Management Jurisdiction
1	Colorado River floodplain lake	2,653 m	lodgepole pine stand	constructed or modified	Rocky Mountain National Park
2	Chickaree Lake	2,828 m	yellow lilies around lake, surrounded by lodgepole pine stand	accessibility concerns	Rocky Mountain National Park
3	Bighorn Lake and neighboring ponds	3,345–3,376 m	near treeline, Engelmann spruce and subalpine fir	accessibility concerns, stratigraphy concerns	Rocky Mountain National Park
4	Unnamed lake near Onahu Creek	2,721 m	lodgepole pine stand, riparian montane community	rocky bottom	Rocky Mountain National Park
5	Onahu Creek fen	2,695 m	Riparian montane community, surrounded by a lodgepole pine stand	within the influence of Onahu Creek	Rocky Mountain National Park
6	Columbine Lake	2,627 m	Colorado blue spruce; residential landscaping	accessibility concerns, modified	Private Homeowner's Association
7	Kettle ponds in Grand Lake Metropolitan Recreation District	2,633–2,644 m	sedges and yellow lilies around and in lakes, lodgepole pine stand	unmodified, smaller bodies dry out	City of Grand Lake

8	Whiskey Fen	2,792 m	sedges, lodgepole pine and Rocky Mountain juniper	Unmodified, 2+ m probe penetration	Arapahoe National Forest
9	Pancake Fen	2,835 m	sedges, lodgepole pine, Douglas fir, and Colorado blue spruce	Frequent human use, <1 m probe penetration	Arapahoe National Forest
10	Hatchet Fen	2,930 m	sedges, lodgepole pine, Douglas fir, and Engelmann spruce	Unmodified, tree hummocks present, >1 m probe penetration	Arapahoe National Forest