

**Using diatoms preserved in peat deposits to track environmental changes in the Cordillera  
Vilcanota, Peru**

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## Abstract

Recent climate change in alpine regions is outpacing many other parts of the world, threatening to degrade specialized ecosystems and their associated ecological services. Long-term data are sparse, especially in remote regions such as the Cordillera Vilcanota in Peru. Diatoms preserved in peat deposits (known locally as bofedales) have the potential to provide a long-term context for environmental changes resulting from recent climate change. This thesis examined diatom assemblages and other siliceous indicators preserved in a 90 cm peat core to reveal long-term environmental changes spanning the past several centuries. The diatom assemblages indicate that cool, oligotrophic conditions occurred contemporaneously with peat initiation at the end of the Little Ice Age, and that these conditions persisted until the late 20<sup>th</sup> century, at which point drier conditions became prevalent. Recently, the diatoms indicate a shift to wetter and more acidic conditions that are characteristic of a poor fen. These changes may reflect glacial retreat due to climate change or ENSO events, the latter of which is known to modulate precipitation in the western tropical Andes. This research also tracked peat accumulation rates that are amongst the highest in the world, highlighting the important role these environments play in carbon sequestration. Although there have been few paleoenvironmental studies in this region, these data generally match interpretations made from other archives such as ice cores and high alpine lake sediments. This thesis is the first to provide the environmental history of a bofedal in the Cordillera Vilcanota using diatoms and demonstrates the potential for future research in this region using siliceous bioindicators preserved in peat deposits.

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## Table of Contents

Abstract.....	2
Acknowledgements .....	3
List of Figures.....	5
Supplementary Figures .....	5
List of Tables .....	5
Introduction and Literature Review.....	5
General Introduction .....	6
Cordillera Vilcanota, Puna.....	8
Bofedales .....	10
Paleoenvironmental Approaches .....	13
Materials and Methods .....	18
Study Site Description .....	18
Core Collection and Chronology .....	18
Diatom Sample Preparation and Enumeration.....	20
Data Analysis .....	21
Results .....	22
Core Chronology.....	22
Diatom Assemblages .....	25
Discussion.....	28
Chronology .....	28
Diatom Assemblages .....	30
Conclusions.....	35
Summary.....	37
References .....	38
Appendix .....	44

## List of Figures

Figure 1: Bofedal study site .....	18
Figure 2: Peat core and coring device .....	19
Figure 3: CRS age-depth models.....	22
Figure 4: $^{14}\text{C}$ model .....	23
Figure 5: Estimated peat accumulation rate .....	25
Figure 6: Core stratigraphy.....	28

## List of Tables

Table 1: $^{14}\text{C}$ dating data.....	24
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## Supplementary Figures

Figure S1: Clam $^{14}\text{C}$ model for the lower bofedal core.....	44
Figure S2: Common diatom taxa observed in the bofedal core .....	45

## Introduction and Literature Review

### *General Introduction*

Global temperatures have risen  $\sim 1^{\circ}\text{C}$  since preindustrial times (IPCC, 2018). This warming is closely linked with anthropogenic sources of greenhouse gases such as carbon dioxide, nitrous oxide, and methane, which trap heat within the atmosphere (IPCC, 2014). Notable effects of recent warming include Arctic sea ice decline (Kwok, 2018), ocean heat content increase (Roemmich et al., 2015), and extremes in precipitation (Donat et al., 2016). However, at local and regional scales, the impacts of climate change vary. For example, precipitation is expected to increase at high latitudes but decrease in regions including Central and South America, the Caribbean, the Mediterranean, and Indonesia (Polade et al., 2014). This variability necessitates that regional studies are undertaken to provide local management with the ability to construct science-based policy.

Of particular importance to climate change studies are regions where warming outpaces the global average. Mountainous regions are one area where temperature increases occur at a higher rate relative to lowland areas (Bradley et al., 2006). This phenomenon, often referred to as elevation dependant warming (EDW), has been observed globally (Wang et al., 2013). Even in regions where cooling is occurring at low elevations, temperatures have continued to increase at high elevation (Vuille et al., 2015). Several mechanisms are thought to contribute to EDW (Mountain Research Initiative, 2015). Chief among these is albedo; as snow and ice (with high reflectivity) melt at high elevations, more bare earth and unfrozen water (with relatively low reflectivity) is exposed, increasing the amount of incoming radiation that is absorbed. Other mechanisms such as increased water vapour and a rising condensation level have been proposed as further contributing factors, but require further study (Mountain Research Initiative, 2015).

EDW is thought to be playing an important role in the tropical Andes, where temperatures are rising at a rate double the global average (Vuille et al., 2003).

Mountain regions act as “water towers” because they play an essential role in water storage via glaciers, lakes, and wetlands. An important “water tower” in southeastern Peru is the Cordillera Vilcanota (13° 59' S; 70° 56' W), which is the second most glacierized region in the country and contains the largest tropical glacier on the planet (Quelccaya Ica Cap). This region experiences a pronounced wet season in the austral summer due to prevailing easterlies, which direct moisture-rich air from the Amazon basin into the mountains (Vuille, 2013). However, throughout the rest of the year, precipitation is low. With recent warming, glacial cover in the Cordillera Vilcanota has declined 35% between 1985 and 2010 (Hanshaw and Bookhagen, 2014). This mountain range supplies water to millions of people downstream for domestic, agricultural, and hydropower purposes, and continued declines in glacier volume will challenge its ability to supply water in the future (Bradley, 2006). Moreover, contemporaneous increases in population and temperature are likely to exacerbate water scarcity (Viviroli et al., 2007). In the Peruvian Andes, where many people rely on agriculture and pastoralism, water scarcity amplified by glacial retreat has resulted in intercommunity struggles for available water (Orlove et al., 2019). The construction of canals may provide water for one community but deprive it from another (Orlove et al., 2019). Declines in the alpine cryosphere mean that ecosystem services, such as irrigation, drinking water, hydropower, climate regulation, and water storage, are reduced (Mukherji et al., 2019).

An important aspect of the water storage network in the Cordillera Vilcanota are peatlands, known locally as *bofedales*. Bofedales are environments where the rate of organic material accumulation outpaces decomposition. The result is large deposits of organic soil.

Bofedales often form downstream from glaciers where topographical and hydrological conditions favour water pooling (Fonkén, 2015). A year-round water supply promotes the growth of cushion plants, such as *Distichia muscoides*, which dominate the landscape (Salvador et al., 2014). These cushion plants act as “sponges”, absorbing glacial meltwater and releasing it slowly over time (Fonkén, 2015). However, without an abundant supply of water, bofedales are prone to desiccation, which can reduce the water retaining capacity of the region. Increasing aridity, as a product of climate change, is expected to negatively affect bofedales’ ability to regulate water storage (Fonkén, 2015).

#### *Cordillera Vilcanota, Puna*

The puna grassland ecoregion is the most commonly found biome in the Peruvian Andes. It extends southward from Bolivia into northern Argentina and Chile and occurs between the elevations of ~3200 to 6000 m a.s.l. (Salvador et al., 2014). It is characterized by the mountainous topography in which it occurs, and the numerous pastures and high alpine lakes that dot the landscape (WWF, n.d.). At its northern reaches, this biome is characterized by wetter conditions, while aridity increases to the south. Wetter conditions favour the proliferation of grasses such as *Agrostis*, *Calamagrostis*, *Festuca*, *Paspalum* and *Stipa*, in addition to small plants including *Azorella*, *Baccharis*, *Gentiana*, *Geranium*, and *Lupinus* (WWF, n.d.). Where topography results in poor drainage, cushion plants like *Distichia* dominate (Fonkén, 2015). Vegetation in the puna region is thus linked to both climate and physiography.

The climate of the Cordillera Vilcanota is determined by a complex interplay of mesoscale and synoptic processes. Precipitation during the austral summer is enabled by the Bolivian High, a semi-permanent zone of high pressure located at 17° S, 70° W. The Bolivian High allows easterlies to develop as far south as 21° S (Garreaud, 2009). This flow advects

moisture from the Amazonian basin, carrying it towards the Andes. However, most water vapour is transported by surface winds that flow south along the eastern Andes and north along the western slopes due synoptic scale circulation. This circulation is caused by permanent zones of low pressure in continental South America and high pressure in the south Pacific and Atlantic (Garreaud, 2009). These processes are further complicated by the influence of sea surface temperature (SST) and the El Niño Southern Oscillation (ENSO).

Temperature and precipitation variability are strongly modulated by both SST and ENSO. During El Niño years, in which warmer conditions prevail in the equatorial Pacific due to weakening of the trade winds, temperatures in the Andes are notably warmer. Temperatures lag behind SST by approximately one to two months (Vuille et al., 2000). Overall, the region receives reduced precipitation during El Niño events. This is mainly due to changes in upper level wind patterns that inhibit westerly transport of moisture (Vuille et al., 2000). As a result, the eastern part of the cordillera is less affected by variations in SST, but the western part of the region receives significantly decreased precipitation. Together this El Niño-induced variability decreases accumulation and increases ablation of Andean glaciers (Francou et al., 2004). ENSO can therefore contribute to interannual climate variability, in addition to exacerbating glacial declines.

The Cordillera Vilcanota occurs at the transition between wet and dry puna. Mean monthly minimum temperatures are just above freezing and mean monthly maximum temperatures are about 15-16°C (Salzmann et al., 2013). Temperature minimums are most different between summer and winter months (Salzmann et al., 2013); however, diurnal temperature swings are much greater than seasonal differences (Salvador et al., 2014). Mean annual precipitation is just below 700 mm, although variability across the cordillera exists (Perry

et al., 2013). Most of this precipitation falls in the distinct wet season that occurs from December to March (Perry et al., 2013; Salvador et al., 2014). Instrumental data are sparse but indicate that the regional climate has warmed since at least 1965. The two closest meteorological stations to my study site, Ccatcca and Pomacanchi, have both recorded significant temperature increases. At the Ccatcca station, average daily minimum and maximum temperatures were 1.3 and 15.3°C from 1965-2014 (Michelutti et al., 2019a). From 1985-2014, the Pomacanchi station recorded daily minimum and maximum values of 2.8°C and 17.2°C, respectively (Michelutti et al., 2019a). The instrumental record shows that daily minimum temperatures at the Ccatcca station increased by 1.08°C, while they rose by 0.80°C at the Pomacanchi station (Michelutti et al., 2019a). This recent warming has the potential to degrade the ecosystems of the Cordillera Vilcanota.

### *Bofedales*

Bofedales in the Cordillera Vilcanota are characterized by edaphic humidity, peat accumulation, and a year-round green appearance (Fonkén, 2015). Most peatlands in the region are thought to be minerotrophic (receive their water from streams or springs) as opposed to the ombrotrophic bogs (receive all of their water and nutrients from precipitation) common in northern latitudes. One study assessing bofedales along a north-south transect in the Peruvian puna region observed pH ranging from 4.5-7.9, peat thickness from 0.36-2.5 m, and a high degree of organic matter content, especially where *Distichia* dominates (Salvador et al., 2014). The year-round green appearance of bofedales is due to the ability of cushion-forming plants to store water. The water that supplies the bofedales is thought to come from both glacial melt and groundwater recharge. For example, some bofedales are exclusively supplied by groundwater that is annually recharged from precipitation in the austral summer (Cooper et al., 2019).

Groundwater flows downslope where it is stored in pools formed by cushion plants before contributing to stream flow (Mosquera et al., 2016). Yet, other studies also demonstrate that wetlands are intimately linked to glacial discharge. Bofedales have been observed to grow in size in response to increased glacial melt (Dangles et al., 2017; Polk et al., 2017). Glaciers and the precipitation that ultimately refills groundwater are tied to climate, and thus changes over the next century are likely to impact their ability to store water and support life.

Bofedales are biodiversity hotspots in extreme environments. They are primarily dominated by cushion-forming plants such as *Distichia* and *Plantago tubulosa* (Fonkén, 2015). Species of *Distichia*, such as *D. muscoides*, support various fauna by providing rich forage. For this reason, bofedales are home to wild camelids such as Guanaco (*Lama guanicoe*) and Vicuña (*Vicugna vicugna*), in addition to supporting a variety of non-native livestock such as horses, cattle, and sheep (WWF, n.d.). Bofedales also act as gathering spots for avifauna, and often are occupied by species such as Yellow-Billed Teals (*Anas flavirostris*), Andean Geese (*Chloephaga melanoptera*), and the Golden-Spotted Ground Dove (*Metriopelia aymara*), among others (Fonkén, 2015). Bofedales also support rodents, amphibians and larger predators such as Cougars (*Puma concolor*) and Culpeo (*Lycalopex culpaeus*). The rich biodiversity of bofedales is tied to their ability to store water.

When the cushion forming plants of bofedales die, organic material accumulates, causing carbon storage to occur. Bofedales are thought to experience faster accumulation rates than other peatlands around the globe due to a 12-month long growing season (Cooper et al., 2010). Boreal peat bogs at northern latitudes have an approximate growth rate of 0.3-0.6 mm yr<sup>-1</sup> (Loisel and Garneau, 2010). The long-term apparent rate of carbon accumulation (LARCA) for these northern peatlands ranges from 13-30 g C m<sup>-2</sup> yr<sup>-1</sup>. However, the recent rate of carbon

accumulation (RERCA) is higher, with a value of 73-83 g C m<sup>-2</sup> yr<sup>-1</sup> (Loisel and Garneau, 2010). In temperate forested peatlands, accumulation is estimated to be 0.25 mm yr<sup>-1</sup> with a LARCA of 17.5 g C m<sup>-2</sup> yr<sup>-1</sup> (Ott and Chimner, 2016). Lowland tropical peatlands are thought to have some of the fastest accumulation rates globally. Peatlands in Indonesia, for example, are thought to have experienced accumulation rates of over 2 mm yr<sup>-1</sup> with corresponding carbon accumulation rates of over 90 g C m<sup>-2</sup> yr<sup>-1</sup> (Page et al., 2004). Although this accumulation slowed down approximately 8000 years ago, the mean LARCA in Indonesia is still 56 g C m<sup>-2</sup> yr<sup>-1</sup> (Page et al., 2004). Yet, evidence exists that bofedales may have the highest rates of accumulation today. In Ecuador, the LARCA of alpine peatlands is estimated to be 46 g C m<sup>-2</sup> yr<sup>-1</sup> (Chimner and Karberg, 2008). In *Oxychloe* bofedales, peat accumulation was found to be over 2 mm yr<sup>-1</sup> in some peat bodies, with LARCAs ranging from 70-292 g C m<sup>-2</sup> yr<sup>-1</sup> (Earle et al., 2003). Further, RERCA is as high as 1033 g C m<sup>-2</sup> yr<sup>-1</sup>, an order of magnitude greater than boreal peatlands (Earle et al., 2003). A study of a *D. muscoides* bofedal in Colombia found accumulation rates as high as 30 mm yr<sup>-1</sup>. These data suggest that bofedales are the most effective carbon accumulating peatlands in the world today.

Despite the role they can play as carbon sinks, bofedales can also contribute greenhouse gases (GHGs) to the atmosphere when they are degraded. Bofedales can be degraded in many ways, most notably by overgrazing, peat cutting, mining, and miscellaneous infrastructure development (Fonkén, 2015). Although studies are relatively scarce, investigated peatlands show signs of excessive overgrazing (Salvador et al., 2014; Machaca et al., 2018). Grazing is demonstrated to reduce *Distichia* cover (Danet et al., 2017), which can result in peatland disturbance. Further, non-native hoofed animals, such as horses and cattle, cause more disturbance to vegetation and increase erosion more than llamas (Deluca et al., 1998). Overall,

overgrazing contributes to increased GHG emissions (Sánchez et al., 2017). Peat cutting is understudied in the Andes, but research from boreal regions indicate that peat extraction contributes to GHG emissions in addition to causing drainage issues (Holmgren et al., 2008). In areas of Peru where mining and infrastructure development occurs, the impacts on bofedales are severe (Salvador et al., 2014). Most damaging is the construction of roads, which disrupt hydrologic flow (Salvador et al., 2014). A common outcome of disturbance is the draining or drying of peat bodies, which leads to GHG emissions.

Global emissions from drained peatlands are large; they emit approximately 1.3 Gt CO<sub>2</sub> annually, or approximately 5.6% of annual anthropogenic CO<sub>2</sub> emissions (IUCN, 2017). Investigations into the impact of peat drainage in the Andes reveal that bofedales produce more methane and carbon dioxide emissions when the water table is lowered (Planas-Clarke et al., 2020). This is likely due to increases in bacterial and fungal biomass that accompanies drainage, and the associated increases in organic matter decomposition (Jaatinen et al., 2008). Drainage can increase soil temperature, which may in turn force bofedales to become net emitters (Gómez et al., 2008). The relationship between water drainage and GHG emissions is well established; it implies that climate change induced drying of bofedales could create a positive feedback loop that produces further emissions and reduces the water storage ability of these environments.

### *Paleoenvironmental Approaches*

Despite the potential impacts of recent climate change, little is known about the long-term environmental history of bofedales. Assessing the impacts of climate change in the Cordillera Vilcanota is difficult. Instrumental data are sparse, and it is difficult for satellite data to be validated (Bradley et al., 2006). Moreover, mountains are challenging environments in

which to execute fieldwork. Bofedales are subject to changes that take place over long temporal scales. To understand changes resulting from climate warming in the present, it is thus necessary to acquire an environmental context that can be used to inform ongoing research and management. The use of paleoenvironmental archives is one such technique that is increasingly being used to understand environmental change.

Paleoenvironmental archives allow for the reconstruction of past environmental conditions. Such archives include lake and ocean sediments, tree rings, ice cores and geological formations. However, available archives are limited to the environment in which they form. For example, it is not possible to use dendrochronology above the treeline to track past changes. Paleolimnology is the study of physical, chemical, and biological indicators preserved in lake sediments used to infer past environmental conditions (Smol, 2008). Since peatlands accumulate over time, similar to lake sediments, they make useful historical archives because buried plant remains contain physical, chemical, and biological indicators of past change. For example, Rühland et al. (2006) used paleolimnological approaches in a peat core to show climatic changes in the Himalayas. In the Cordillera Vilcanota, available paleoenvironmental archives include ice cores, lake sediments, and peat cores.

Ice core records from the Quelcaya Ice Cap (QIC) in the Cordillera Vilcanota reveal an annually resolved 1800-year-old record of environmental change (Thompson et al., 2013). Prior to ca. 1100 CE, conditions in the Cordillera Vilcanota were relatively stable. The Medieval Climate Anomaly (MCA), which took place from 1100-1300 CE, corresponded to a decrease in net accumulation on the QIC (Thompson et al., 2013). At the onset of Little Ice Age (LIA) in 1520 CE, accumulation increased until it peaked in 1680, before declining to record lows towards the end of the LIA in 1800 CE (Thompson et al., 2013). From 1800 CE onwards, there

has been a gradual increase in accumulation on the QIC (Thompson et al., 2013) Increases in net accumulation are linked to moister climates whilst lower accumulation values indicated increased aridity. Thompson et al. (2013) therefore attribute changes in accumulation on the QIC to variations in regional precipitation. They note that in the last century,  $\delta^{18}\text{O}$  enrichment has occurred, suggesting a warmer SST. This, coupled with glacial retreat that is unprecedented in at least the last 5,000 years (Thompson et al., 2006), illustrate that climate change is already impacting the region.

The multitude of high alpine lakes in the Peruvian Andes corresponds to a wide availability of lake core records. A 24,700-year-old record from Lake Pacucha ( $13.61^{\circ}\text{S}$   $73.31^{\circ}\text{W}$ ), approximately 250 km away from my study site, indicates that deglaciation began approximately 23,000 yr. BP, resulting in declining lake levels until ca. 10,000 yr. BP (Hilleyer et al., 2009). Lake level gradually normalized, only declining again at ca. 750 yr. BP, during the middle of the MCA (Hilleyer et al., 2009). It is generally thought that from 12,000-8000 yr. BP, arid conditions dominated, followed by a period of wetter conditions until 4000 yr. BP, at which point both warm and wet conditions prevailed (Stansell et al., 2013). More recently, paleolimnological investigations of sediment recovered from an Incan pot submerged in Laguna Sibinacocha (the largest lake in the Cordillera Vilcanota) reveal that lake levels rose in response to wetter conditions during the LIA and that these higher levels have persisted for the past few hundred years (Michelutti et al., 2019b). Bioindicators preserved in Peruvian lake sediments reveal increased thermal stratification, as a result of climate warming, is occurring (Michelutti et al., 2015). Overall, Holocene lake cores provide significant paleoclimatic context and more recent lake cores are in agreement with the QIC record (Michelutti et al., 2020)

Despite their importance, little is known about the long-term environmental history of bofedales in the Cordillera Vilcanota and elsewhere in the Andes. Schitteck et al. (2016) examined an Argentinian (24°S) peatland profile and determined that dry periods occurred from 2100 cal. yr. BP to 1800 yr. BP and again from 1300-1150 cal. yr. BP and 950-850 cal. yr. BP. The last dry period corresponds to the onset of the MCA. It was also determined that wetter conditions dominated at the end of the LIA, a finding that is supported by other records (Thompson et al., 2013; Kock et al., 2019a; Kock et al., 2019b). In a study from the same peatland, oxygen isotopes revealed that arid conditions persisted from 2190-2120, 1750-1590, 1200-1080, and since 130 cal. yr. BP (Kock et al., 2019a). In contrast, humid conditions occurred from 2750-2250 and from 600-130 cal. yr. BP (Kock et al., 2019a). The Lagunillas peatland in Chile (27°S) revealed that humid conditions were prevalent from 1530-1270 cal. yr. BP, before giving way to more arid conditions associated with the MCA (Kock et al., 2019b). A similar increase in precipitation that peaked around 150 cal. yr. BP was also observed, and drier conditions are apparent after -30 yr. BP (Kock et al., 2019b). Only one Peruvian bofedal (14°S) record is available, although it is located approximately 400 km away from my study site. It indicates that stable humid conditions persisted from 1800-1200 cal. yr. BP, followed by a dry period from 1200-750 cal. yr. BP, after which the record is unstable, possibly due to decreased peatland stability during the LIA (Schitteck et al., 2015). Together these studies illustrate the limited research in the region; they also confirm that Peruvian bofedales are understudied despite their importance to regional hydrology. Moreover, existing peat studies in the Andes have so far been limited to  $\delta^{18}\text{O}$ , element ratios, and pollen. In this study, I will use algal bioindicators for the first time in Andean peat core studies to capture ecological changes that occur in response to climate variability.

Diatoms (*Bacillariophyceae*) are well-established aquatic bioindicators used for tracking a wide array of environmental changes (Smol and Stoermer 2010). They are microscopic, siliceous algae that are distributed globally. Diatoms are found in lakes, rivers, the ocean, soil, wetlands, and even in glacial ice. Their wide distribution, sensitivity to environmental change, and the fact their siliceous cell walls preserve well, makes them valuable archives of environmental change (Smol and Stoermer, 2010). Diatom distribution is influenced by a variety of physical factors, including temperature, the turbulence of the water column, ice cover, and light (Battarbee et al. 2001). Species composition is also highly dependent on nutrient levels, salinity, and pH (Battarbee et al. 2001). This sensitivity to different abiotic factors is what makes diatoms useful environmental indicators. Diatom analysis has been used to study a variety of environmental concerns such as lake acidification, eutrophication, drought, and climate change.

Given recent climate change and the evidence of accelerated warming in the Andes, it is crucial that studies be undertaken to provide a paleoclimatic context for recent changes. The bofedales of the Cordillera Vilcanota are largely unstudied despite their ecological and societal importance. These bofedales are biodiversity hotspots, valuable carbon sinks, and play a critical role in regulating water flow from glaciers and groundwater. Increased aridity threatens to dry them out, a consequence that would likely reduce their ability to provide these ecosystem services. In this study, I provide the first record of fossil diatom assemblages from an Andean bofedal core spanning the past several centuries. My research has important implications to environmental managers and policy makers as the data will inform about how rising temperatures and changes in water availability are affecting critical peatland ecosystems in the Andes.

## Materials and Methods

### *Study Site Description*

The bofedal used in this study is located in the Pitumarca district of Canchis province of Peru ( $13^{\circ} 48' 42.9''\text{S}$   $71^{\circ} 06' 15.0''\text{W}$ ; Figure 1). It is located at just over 5000 m a.s.l. and 3.2 km to the west of Lake Sibinacocha. Two small glaciers (areas 2.86 and 1.00 km<sup>2</sup>) are located to the east and north.



Figure 1: Bofedal study site where the study core was extracted. The glacier draining into the bofedale is visible in the background. (Photo taken by Neal Michelutti).

### *Core Collection and Chronology*

A 90 cm peat core was extracted on August 12<sup>th</sup>, 2018 using a peat corer (Figure 2). Glacial clays at the bottom of the core indicate the entire history of peat accumulation was

collected. The core was sectioned into 2 cm intervals in the field, using a sharp blade. Samples were stored in Whirl-pak® bags in cool and dark conditions until being transferred to the PEARL lab at Queen's University. At the lab, samples were freeze dried for further analysis.

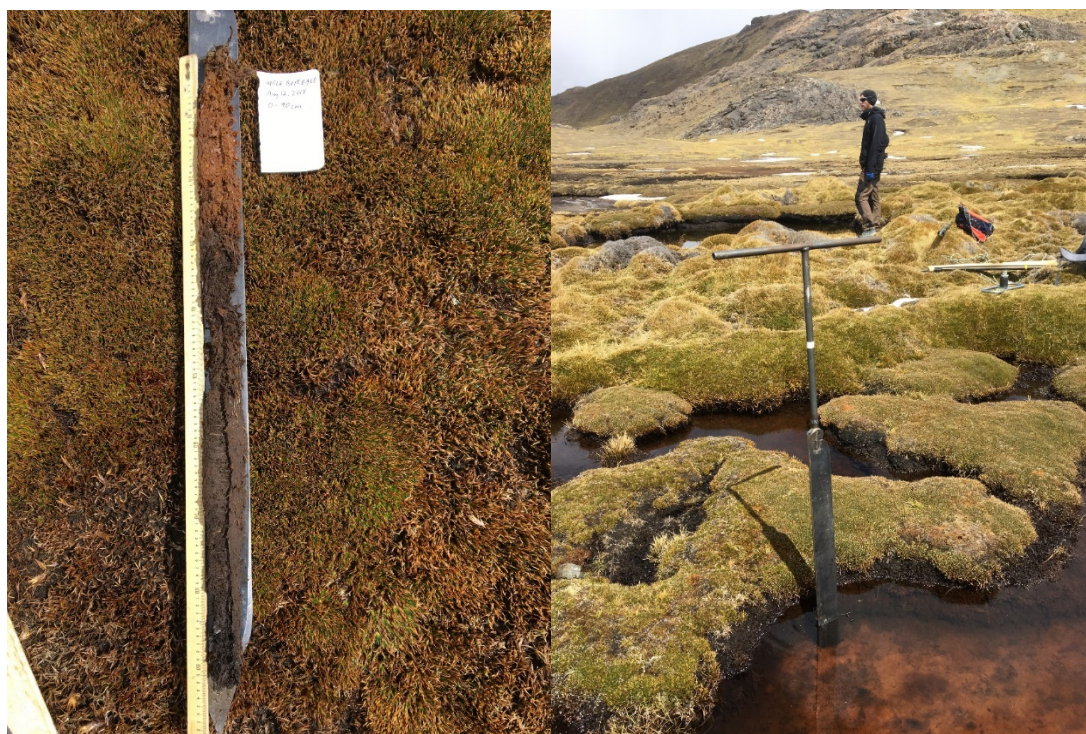


Figure 2: Left: the 90 cm peat core extracted from the upper bofedal site. Right: the peat corer used to extract the bofedal core. (Photo taken by Neal Michelutti).

The uppermost 10-cm of the core was used to establish a  $^{210}\text{Pb}$  chronology using the methods outlined by Appleby et al. (2001).  $^{210}\text{Pb}$  is a short lived (half-life = 22.3 years) radioisotope that can be used to determine the age of a core for the past ~ 150 years (Appleby and Oldfield, 1978). A mortar and pestle were used to breakdown selected samples and approximately 0.5-0.9 grams of peat were then placed in a plastic vial. A silicone septum and 2-ton epoxy were used to seal the samples, which were left to dry for several weeks.  $^{214}\text{Pb}$  was used to infer supported or background levels to which the  $^{210}\text{Pb}$  activity could be compared.

Using a constant rate of supply model (CRS), a  $^{210}\text{Pb}$  chronology was developed with the ScienTissiME package in MatLab<sup>®</sup>.

Radiocarbon dating was used to establish a long-term age profile for the core. Ten samples of non-root material or bulk peat were selected from seven intervals and sent to the radiocarbon laboratory at the A.E. Lalonde AMS Laboratory at University of Ottawa for analysis (Table 1). To corroborate the radiocarbon model, peat samples from a bofedal downslope of the sight were also dated. A radiocarbon model was established using the *Clam* package in R (Blaauw, 2020).

#### *Diatom Sample Preparation and Enumeration*

Diatom sample preparation was accomplished following techniques outlined by Battarbee et al. (2001). Approximately 0.03 grams of peat were measured from 45 samples (sectioned at 2 cm) and placed in a glass scintillation vial. To digest organic material, 15 mL of a 50:50 molar mixture of sulphuric and nitric acid was added to the vials, which were then placed in a ~80°C water bath. Slurries were stirred every 20 minutes for at least two hours. Following the bath, samples were left to cool for 24 hours in order to let siliciclastic material in the slurry to settle. The slurries were then aspirated to remove the acid and deionized water was added; this process was repeated until the pH became neutral. Slurries were tested for neutrality using litmus paper. After this, the neutral slurries were shaken inside the scintillation vial and a pipette was used to extract integrated samples of varying dilutions. Samples were placed on a cover slip and left to dry for 24 hours using a slide warmer. Finally, cover slips were mounted onto slides using Naphrax<sup>®</sup>.

Diatom valves were enumerated using a Leica® DMRB light microscope fitted with differential interference optics at 1000X magnification. Efforts were made to count a minimum of 300 valves per interval (every 2-cm), but where diatoms valves were sparse, a minimum of at least 150 valves was enumerated. Intervals where 150 valves could not be counted were excluded from analysis. Diatoms were identified to the species level where possible using several sources including Krammer and Lange-Bertalot (1986) and Kulikovsky et al. (2010). The ratio of diatoms to other siliceous indicators was calculated using methods outlined by Smol (1985).

### *Data Analysis*

In order to determine assemblage zones of significance, a constrained incremental sum of squares (CONISS) was performed (Grimm, 1987). First, diatom data was converted to relative abundance. Taxa that did not exceed more than 2% relative abundance in two intervals were excluded from analysis. Data was loaded into R and a CONISS analysis was carried out using the *rioja* package (Juggins, 2017). Using the same package, a broken stick analysis was then performed to determine the number of significant zones.

To illustrate the profile, diatoms with a relative abundance of 5% or greater were plotted using C2. The significant zones determined by the CONISS analysis were overlaid on the stratigraphy and the ratios of diatoms to other siliceous indicators including phytoliths, chrysophyte cysts, and protozoan plates, were calculated (Smol 1985) and plotted beside the diatom assemblages.

## Results

### Core Chronology

Unsupported  $^{210}\text{Pb}$  activity is relatively low and declines from approximately  $290 \text{ Bq kg}^{-1}$  before reaching supported levels at 7 cm (Figure 3). These values are about three times lower than surface values recorded in nearby lake sediment cores (Michelutti et al. 2020). There is no peak in  $^{137}\text{Cs}$ , and thus this anthropogenic radionuclide cannot be used to independently corroborate the modeled  $^{210}\text{Pb}$  dates. Although the unsupported  $^{210}\text{Pb}$  activity shows a steady decline to background, the chronology should be interpreted with caution due to such low activity levels.

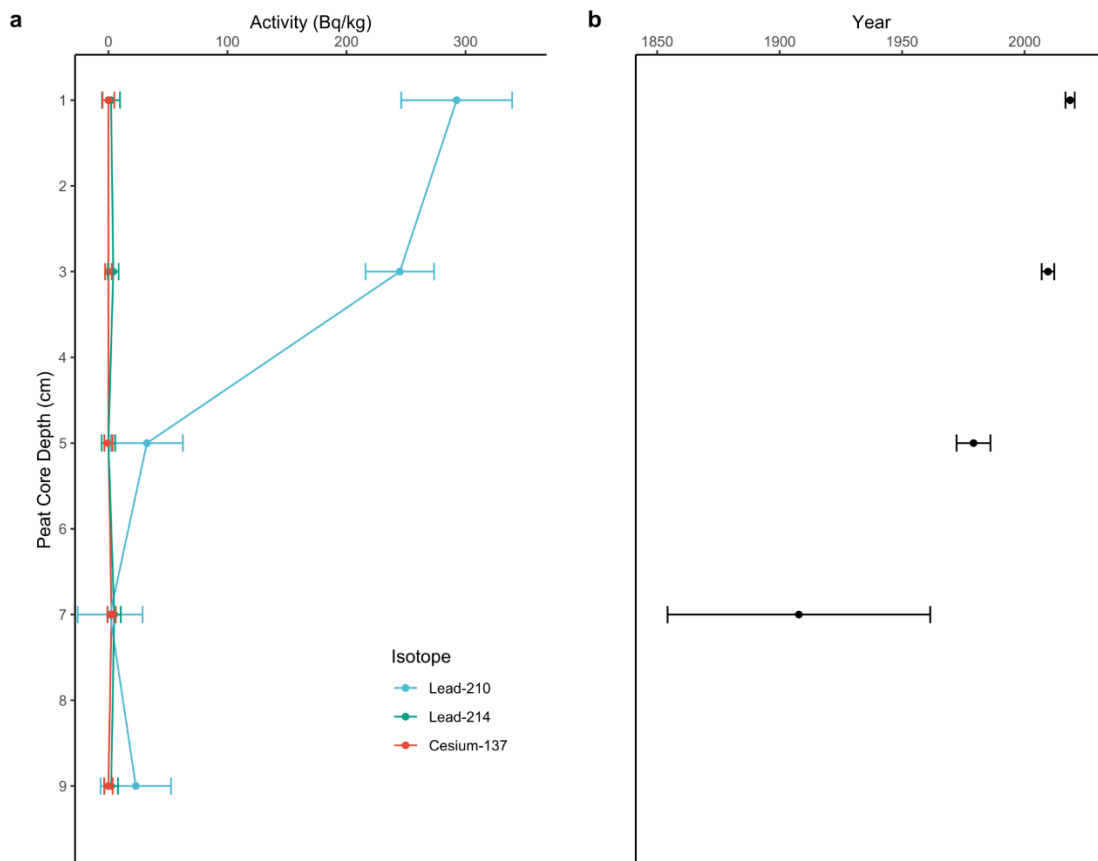


Figure 3: CRS age-depth models using unsupported  $^{210}\text{Pb}$  activities. **a:** Activity of  $^{210}\text{Pb}$ ,  $^{214}\text{Pb}$ , and  $^{137}\text{Cs}$  from the first 10 cm of the peat core. **b:**  $^{210}\text{Pb}$  inferred dates using the CRS model.

$^{14}\text{C}$  dating indicates that the basal age of the core is approximately 227 cal. yr. BP (Figure 4). Towards the base of the core, the radiocarbon ages intersect with the  $^{14}\text{C}$  calibration curve, and thus these ages should be interpreted with caution. Subsequent  $^{14}\text{C}$  dating of a peat core extracted downslope reveals a dating profile similar to the upper bofedal, lending support that  $^{14}\text{C}$  ages are recording the actual dates of peat accumulation (Figure S1). The results of the radiocarbon dating are summarized in table 1.

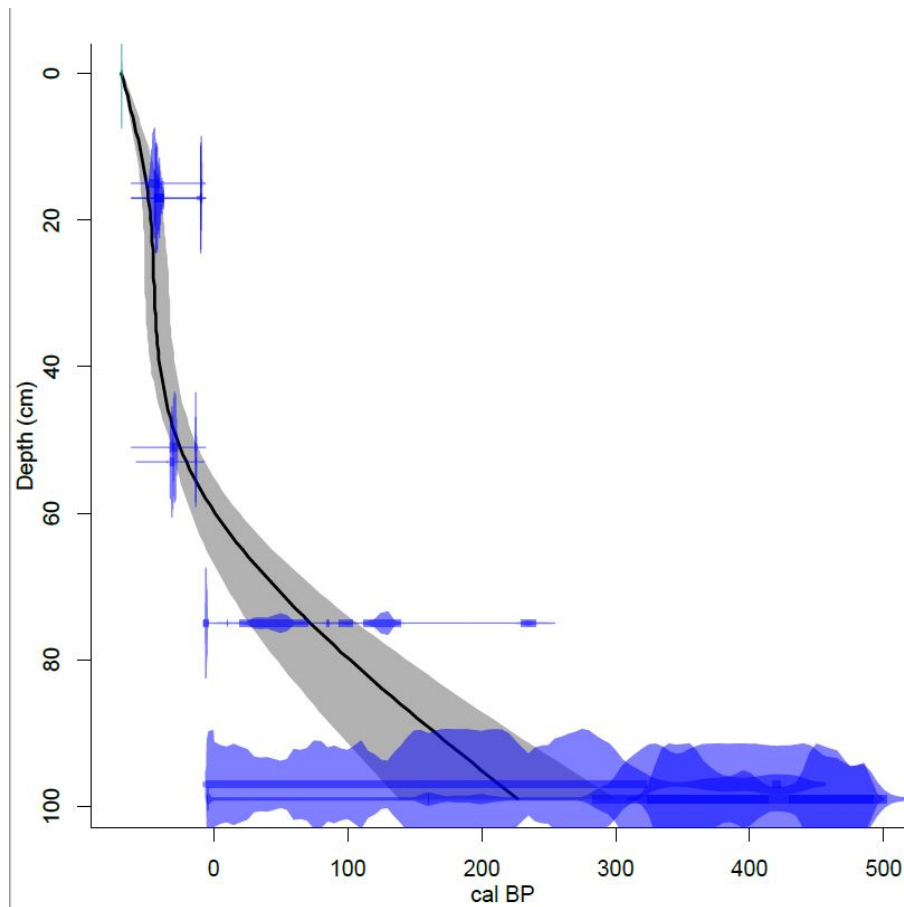


Figure 4: Radiocarbon age model developed using *clam*.

Core	Interval (cm)	Laboratory ID	F <sup>14</sup> C ( $\pm 1\sigma$ )	<sup>14</sup> C yr. BP	$\pm 1\sigma$	Calibrated Age (cal. yr. BP)	Calibrated Age Range (cal. yr. BP)	Material
Ubof 1	14-16	UOC-10804	1.1283 $\pm$ 0.0064	>Modern	45	-50.05	-54.22 to -41.98	Picked peat fascicles
Ubof 1	16-18	UOC-11410	1.1516 $\pm$ 0.0032	>Modern	22	-48.54	-53.11 to -39.87	Picked peat fascicles
Ubof 1	16-18	UOC-11411	1.1540 $\pm$ 0.0032	>Modern	22	-48.54	-53.11 to -39.87	Bulk peat
Ubof 1	50-52	UOC-10805	1.3125 $\pm$ 0.0095	>Modern	58	-24.93	-30.59 to -12.06	Picked peat fascicles
Ubof 1	52-54	UOC-11412	1.2739 $\pm$ 0.0036	>Modern	23	-20.27	-27.85 to -6.9	Bulk peat
Ubof 1	74-76	UOC-12495	0.9915 $\pm$ 0.0028	68	23	72.21	28.03 to 105.89	Bulk sediment, visible organics
Ubof 2	66-68	UOC-10806	1.0881 $\pm$ 0.0080	>Modern	59	212.91	131.65 to 280.63	Picked peat fascicles
Ubof 2	66-68	UOC-10807	0.9741 $\pm$ 0.0084	211	69	212.91	131.65 to 280.63	Bulk peat
Ubof 2	68-70	UOC-11413	0.9524 $\pm$ 0.0028	392	24	226.59	141.63 to 297.94	Bulk peat
Ubof 2	68-70	UOC-11414	0.9562 $\pm$ 0.0069	360	58	226.59	141.63 to 297.94	Picked peat fascicles

Table 1: <sup>14</sup>C dating data and results. The calibrated age range is  $\pm 2\sigma$ .

The radiocarbon model reveals substantial changes in accumulation rate (Figure 5). From the base of the core to the 52-54 cm interval, rates slowly increase from 1.5 to 3.8 mm yr<sup>-1</sup>. At around the 52-54 cm interval, the accumulation rate accelerates rapidly, reaching a peak of 48 mm yr<sup>-1</sup> at the 26-28 cm interval. From this point, accumulation rates decline to the top of the core, falling to a rate of 7 mm yr<sup>-1</sup>.

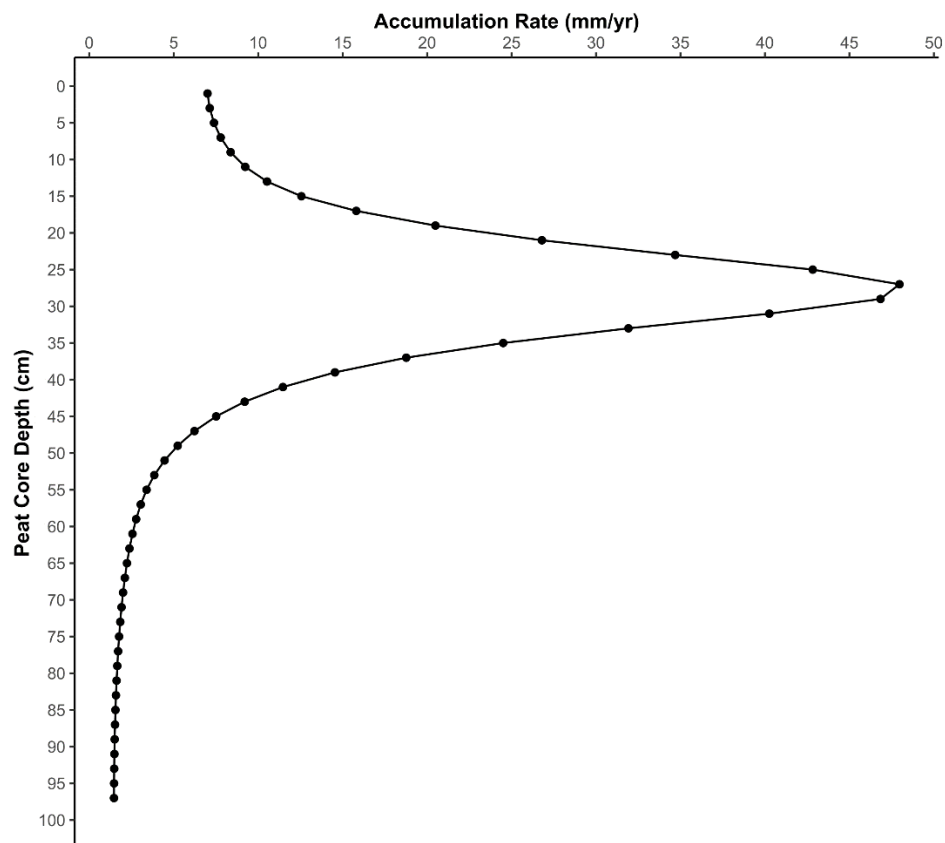


Figure 5: Estimated peat accumulation rate for the upper bofedal core.

### *Diatom Assemblages*

A total of 147 taxa across 42 genera were identified in the core. Discussion will be limited to dominant taxa defined here as those species making up 5% or more of the assemblage in any interval. This criterion was met by 26 taxa, with the most abundant species being *Eunotia arcus*, *Hantzschia amphioxys*, *Nitzschia paleacea*, *Kobayasiella subtileissima*, *Pinnularia* spp. cf. *kuetzingii*, *Sellaphora* spp cf. *subnympharum*, *Staurosira construens*, *Encyonema norvegicum*, and *Staurosirella pinnata* (Figure S2). The CONISS analysis determined there are six significant assemblage zones in the core.

### Zone 1

The first diatom zone extends from 90 cm to 64 cm. Here, there is relatively more diversity in major species compared to the top 35 cm (Figure 6). *S. construens* is dominant from 90 cm to 70 cm. Other benthic taxa such as *E. norvegicum*, *E. falaisensis*, and *E. silesiacum* are present, although in smaller relative abundances. *S. pinnata* appears at the 78-80 cm level and remains prominent in the zone.

### Zone 2

Zone 2 extends from 64 cm to 32 cm. *S. construens* declines abruptly at the start of this zone, before disappearing from the record (Figure 6). The decline of *S. construens* occurs contemporaneously with an increase in the relative abundance of *E. norvegicum*, which reaches a relative abundance of 25% by 58 cm. Following this, there is a small peak in *E. falaisensis* until the 50-52 cm interval. From 50 cm to 42 cm there is an increase in *Ulnaria ulna* that also coincides with an increase in *E. silesiacum*. These increases occur concurrently with relative decreases in *E. norvegicum* and *S. pinnata*. Towards the end zone two, most benthic taxa decline in relative abundance; several *Encyonema* taxa disappear, with the exception of *E. norvegicum*. *P. cf. kuetzingii* increases in abundance from 38-32 cm. At the same time, the number of protozoan plates, chrysophycean cysts, and phytoliths increase relative to the amount of diatom valves.

### Zone 3

Zone 3, which spans from 32-28 cm, contains an abrupt assemblage shift with *S. cf. subnympharum* increasing rapidly from zero values to a relative abundance of 74% (Figure 6). At the 28-30 cm interval *E. norvegicum* increases. Both of the aforementioned taxa decline

rapidly at the end of this zone. The relative number of diatom valves to chrysophycean cysts increases in this zone.

#### *Zone 4*

Zone 4 spans from 28-22 cm. *N. paleacea* is the most dominant taxon in this zone (Figure 6). *H. amphioxys* and *P. cf. kuetzingii* are also prominent. *S. cf. subnympharum* declines as the ratio of diatoms to siliceous indicators continues to decrease (diatoms become less abundant relative to other microfossils).

#### *Zone 5*

Zone 5 occurs from 22-14 cm. Most taxa that have been observed in lower relative abundances disappear. *H. amphioxys* becomes overwhelmingly dominant, followed by *N. paleacea* (Figure 6). From 22 to 16 cm, the relative abundance of diatom valves increases in relation to plates, cysts, and phytoliths.

#### *Zone 6*

At the 14-16 cm interval, *P. cf. kuetzingii* begins to increase as *N. paleacea* and *H. amphioxys* decline (Figure 6). As this transition takes place, the ratio of diatoms to siliceous indicators decline again. At the 12-14 cm interval, *E. arculus* becomes a dominant taxon. From 10 cm to 2 cm, the combined relative abundance of *P. cf. kuetzingii* and *E. arculus* are over 50%. At 4 cm, *K. subtileissima* begins to increase, and becomes the dominant taxon at the surface.

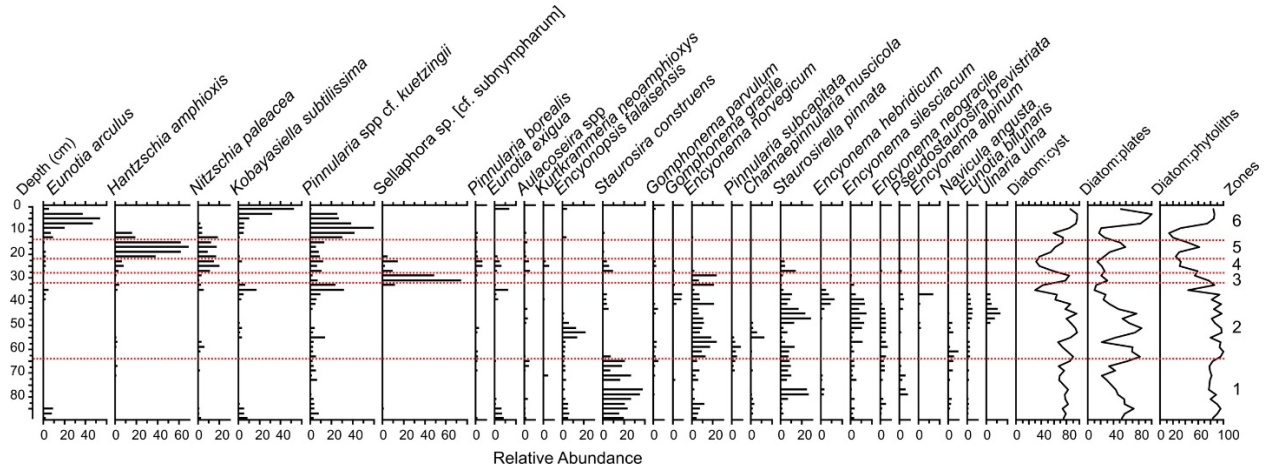


Figure 6: Subfossil diatom assemblage from the bofedal core. Only species with relative abundances above 5% are shown. The ratios of diatoms to chrysophycean cysts, protozoan plates, and phytoliths are shown on the right side of the plot. The stratigraphy is divided into six zones using CONISS; significant zones were determined through broken stick analysis.

## Discussion

### Chronology

As in many high latitude regions (Wolfe et al 2004), dating in high altitude regions can be difficult. The CRS-derived  $^{210}\text{Pb}$  dates indicate the peat is older than what the numerous  $^{14}\text{C}$  ages indicate (Figures 3 and 4). Given the low  $^{210}\text{Pb}$  activities recorded in the peat core, and our inability to corroborate these dates with independent radionuclides such as  $^{137}\text{Cs}$ , the  $^{210}\text{Pb}$  chronology should be viewed with caution. Although the age model developed exclusively from  $^{14}\text{C}$  ages indicate a high rate of peat accumulation (Fig. 5), these values of up to  $48 \text{ mm yr}^{-1}$  are not unprecedented for Andean peatlands (Benavides et al., 2013). In addition,  $^{14}\text{C}$  dates from a nearby bofedal that is slightly lower in elevation record similar age-depth values (Figure S1). For these reasons, the age model derived exclusively from  $^{14}\text{C}$  ages (Fig. 4) is used as the primary chronology.

The radiocarbon model suggests that the entire core is aged 227 cal. yr. BP, with the 90 cm interval being aged 159 cal. yr. BP. Typically,  $^{14}\text{C}$  is not used to date younger sediment or peat due to the Suess effect, which is a term that signifies the decrease in  $^{14}\text{C}$  in atmospheric  $\text{CO}_2$  owing to admixture of  $\text{CO}_2$  produced by the combustion of fossil fuels. Therefore, the carbon model developed for this study is not able to provide yearly approximations for samples that were classified as modern (>200 years old). However, due to the fact that repeated  $^{14}\text{C}$  testing revealed similar dates, the radiocarbon dating model likely provides the best estimate of the core age. Furthermore, this finding can be corroborated. A second peat core was also analyzed and revealed a similar  $^{14}\text{C}$  curve with a comparable acceleration in the accumulation rate and a basal age of approximately 500 yr. BP. In addition, bofedales dominated by *D. muscoides* have been successfully dated with  $^{14}\text{C}$  before (e.g. Hribljan et al., 2015). Overall, the radiocarbon model suggests very fast peat growth.

The radiocarbon dating model calculates a minimum basal age of 141 cal. yr. BP and maximum basal age of ~ 298 cal. yr. BP (95% confidence interval), representing an average peat accumulation rate in the range of 2.7-4.8 mm yr<sup>-1</sup>. This rate varied through time, with recent accumulation rates reaching a maximum of almost 50 mm yr<sup>-1</sup>. These accumulations are large but not unprecedented. Deglaciation during the Holocene thermal maximum is thought to have resulted in accelerated peat expansion in Alaska (Jones and Yu, 2010). More recently, Earle et al. (2003) reported fast accumulation rates in *Oxychloe* peatlands with RERCA values over 1000 g C m<sup>-2</sup> yr<sup>-1</sup>. In Colombia, *Distichia* bofedales were observed to have accumulation rates as high as 30 mm yr<sup>-1</sup> (Benavides et al., 2013). These accelerated rates of peat growth are thought to be due to the onset of favourable conditions, such as increased water supply due to glacial melt (Benavides et al., 2013). Such high peat accumulation rates also hint at the potential of bofedales

to act as effective carbon sinks. Recent deglaciation is a product of climate change and thus it is possible this trend of accumulation could reverse with continued warming once Andean glaciers begin to disappear.

### *Diatom Assemblages*

#### *Zone 1 (90-64 cm) Approximately 159 to 22 cal. yr. BP*

*S. construens* is the taxon in highest relative abundance throughout this zone. *S. construens* is a benthic, typically epilithic diatom commonly found in cold, alkaline lakes (Michelutti et al., 2003; Michelutti et al., 2007). These diatoms are thought to be pioneering taxa that are among the first to colonize new aquatic systems (Lotter and Bigler, 2000). Moreover, they can often reflect prolonged ice cover (Douglas and Smol, 2010). At Lake Sibinacocha, just 3.2 km to the west of the coring site, the *Fragilaria sensu lato* complex (includes *S. construens* and *S. pinnata*) has dominated the assemblage for the last ~ 500 years (Michelutti et al., 2019b). *S. pinnata*, also common in this zone, indicates similar aquatic characteristics to *S. construens*. Both taxa have been observed to be positively influenced by shallow water conditions in Canadian peatlands (Hargan et al., 2015). The assemblage here indicates alkaline, cold and shallow water conditions in an emerging peat environment.

Interestingly, the onset of peat development for this site is thought have occurred towards the end of the LIA. The  $\delta^{18}\text{O}$  record from the QIC indicates that accumulation peaked between 1630-1680 CE before declining until 1800 CE (Thompson et al., 2013). The diatom record begins when accumulation on the QIC is at a record low, suggesting that as glaciers retreated following the LIA glacial maximum, meltwater provided the moisture necessary for peat development, especially where topography favoured pooling. This peat development mechanism

has been proposed for other parts of the Andes (Earle et al., 2003; Benavides et al., 2013).

Furthermore, this explains the older basal ages of peat cores downslope; high precipitation at the beginning of the LIA could have initiated downslope peat development, whereas increased glaciation delayed it at higher elevations until glaciers retreated following the LIA glacial maximum.

#### *Zone 2 (64-32 cm) Approximately 13 to -43 cal. yr. BP*

The beginning of Zone 2 is distinguished with the abrupt decline of *S. construens*. *S. pinnata* remains prevalent and taxa such as *E. norvegicum* become more dominant. *E. norvegicum* has been observed in cold, alkaline water in tundra and hummocky regions (Bahls et al., 2018). These two dominant taxa suggest water conditions are relatively similar between Zones 1 and 2. Other taxa such as *E. falaisensis*, which are known to occur in small alkaline headwaters in the Rocky Mountains, support this (Bahls, 2013). *U. ulna* has been noted worldwide in variety of alkaline, low conductivity environments including lakes and rivers (Salomoni et al., 2002; Kivrak and Uygur, 2012; Severiano et al., 2012;). The small peak in *U. ulna* may thus represent a short period of time with accelerated glacial runoff or water flow within the bofedal. This agrees well with the QIC record, which reveals that precipitation was higher at the onset of current warming period (Thompson et al., 2013).

At the end of Zone 2, the relative proportions of cysts, plates, and phytoliths decline. Zeeb and Smol (1993) suggest that higher number of cysts are indicative of cool, oligotrophic conditions, which would support the interpretation of diatom assemblage. However, the proportion of cysts declines less dramatically and rebounds quickly relative to plates and phytoliths. High concentrations of plates are thought to indicate woody, poorly decomposed peat (Douglas and Smol, 1987), in addition to drier conditions (Rühland et al., 2000). Likewise,

increased proportions of phytoliths have been shown to be linked to dry conditions in peat studies (Rühland et al., 2000). These siliceous indicators reveal that, although cool, oligotrophic conditions likely persisted for most of zone two, there is an abrupt shift to drier conditions at approximately 35 cm.

The  $^{14}\text{C}$  model suggests that the end of zone two occurs in the late 20<sup>th</sup> century. Anthropogenic warming may have contributed to increased glacial melt in the 20<sup>th</sup> century, producing the cool, oligotrophic conditions inferred from the diatom assemblage in this zone. On the other hand, accumulation on the QIC also indicates that precipitation gradually increased during the 20<sup>th</sup> century (Thompson et al., 2013). Towards the end of zone two, there appears to be a shift to drier conditions at the end of the 20<sup>th</sup> century that is reflected in the diatom assemblage and from proportions of phytoliths and plates.

*Zone 3-5 (32-14 cm) Approximately -44 to -50 cal. yr. BP*

Zone 3 begins with a marked increase of *S. cf. subnympharum*. Little data exist on this taxon and without a positive identification it is difficult to interpret this change. During this period, diatom valves become less abundant in the peat core. This could reflect a shift to more acidic conditions, but the high proportion of well-preserved plates precludes this (Beyens and Meisterfield, 2001). Therefore, the *Sellaphora* spp. may instead be more competitive in the drier conditions that can be inferred from the high proportions of plates and phytoliths that occur concurrently.

*Sellaphora cf. subnympharum* declines quickly as *N. paleacea* and *H. amphioxys* emerge and become more dominant. *N. paleacea* typically indicates pollution, nutrient-enriched waters, or waters with higher conductivity (Holmes and Taylor 2015). The presence of *N. paleacea* is

difficult to explain, but its emergence may be due to short term changes in water chemistry (Woodbridge and Roberts, 2010). *H. amphioxys* is an aerophilic taxa that can be found in high abundance in neutral to alkaline soils (Johansen, 2010). Its high abundance infers drier conditions occurred in the section of the core.

The radiocarbon model suggests that Zones 3-5 represent the late-20<sup>th</sup> century. Dominant taxa such as *H. amphioxys* indicate dry conditions, an interpretation that is supported by high relative proportions of cysts, plates, and phytoliths. During this section in the core, the peat accumulation rate declines quickly, further supporting the idea that dry conditions were prevalent at this time. There are several possible explanations for this shift. First, the marked dry conditions may be due to temperature increases. Temperatures have risen significantly in this region of the Andes since at least 1950 (Vuille et al., 2015). As a result, evaporative losses from the bofedal surface could occur, leading to dry conditions at the surface, even if the underlying water table is maintained. Although it is difficult to say without annually resolved dates, this shift to drier conditions may reflect changes in ENSO. Vuille et al. (2000) has demonstrated that El Niño (La Niña) events produce drier (wetter) conditions over the central Andes. The 1980s and 1990s witnessed four of the five strongest El Niño events on record. In addition, the QIC record indicates that there was a sudden, albeit short-lived, decrease in accumulation at the end of the 20<sup>th</sup> century (Thompson et al., 2013). Since precipitation and temperature in the western Cordillera Vilcanota is modulated by ENSO, it is possible that the diatom assemblage in this section of the core reflects this period.

*Zone 6 (14-0 cm) Approximately -52 to -67 cal. yr. BP*

Zone 6 begins with a notable decline in aerophilic taxa such as *H. amphioxys*. At the same time, the relative abundances of *Pinnularia* cf. *kuetzingii* and *E. arculus* increase. *E. arculus* has been found in small lakes with submerged sedges in Canada (Bahls et al., 2018). Importantly, *E. arculus* is an acidophilic taxon (Luís et al., 2013). It is the first dominant acidophilic taxa in the record, suggesting a shift to more acidic conditions in the bofedal. This could be due to the natural hydrosere development. Gaiser and Rühland (2010) note that *E. arculus* can indicate increasing paludification and the development of a poor fen. This is supported by increases in the relative proportion of diatoms to cysts, plates, and phytoliths, which implies wetter conditions (Rühland et al., 2000; Rühland et al., 2006).

Although aerophilic taxa such as *H. amphioxys* usually indicate late stages of peat development (i.e. raised bogs), the process of hydrosere succession is not necessarily unidirectional and can be influenced by climate (Gaiser and Rühland, 2010). Given the timing of this shift from dry to wetter conditions, it is possible there is climatic influence. Four strong La Niña events occurred from 1998-2011, while there were no significant El Niño events. This means that the bofedal could have experienced increased precipitation during this period (Vuille et al., 2000).

Near the top of the core, there is a shift to *K. subtilissima* and *E. exigua*. The former is a significant indicator of wet fens that are slightly acidic (Hargan et al., 2015). *E. exigua* is a well-known indicator of acidic conditions (Furey, 2012). Together, these two taxa suggest that there has been a shift to wetter, acidic conditions within the bofedal. This may be a result of the aforementioned strong La Niña events impacting the region around this time. It may also be due to accelerated melting of glacial ice. It is not possible to compare the observed assemblage

changes in Zone 6 to the QIC record as it ends in 2003. As climate change progresses, it is possible that wetter conditions may prevail for a period of time before glaciers disappear from the Cordillera Vilcanota entirely.

### *Conclusions*

The  $^{14}\text{C}$  chronology indicates that peat development likely began approximately 230 cal. yr. BP. Since that time, about one metre of peat developed, adding to a growing body of evidence that Andean peatlands have some of the fastest accumulation rates worldwide. The bofedal core shows several diatom assemblage changes that appear to be in response to regional climate controls. Pioneering taxa such as *S. construens* and *S. pinnata* indicate that peat growth began amidst a cool, alkaline, and oligotrophic conditions. This environment may have been linked to glacial retreat and melt following the LIA glacial maximum. Increasing precipitation, as indicated by the QIC record, and climate change-facilitated glacial melt, are likely to have further favoured conditions that led to peat growth and the alkaline, cold-water diatom assemblages observed in the lower zones of the peat core.

Towards the end of the 20<sup>th</sup> century, the diatom assemblage and relative proportions of cysts, plates, and phytoliths indicate that drier conditions became prevalent. Since precipitation in the western tropical Andes is largely modulated by ENSO, this assemblage shift may be attributed to the series of strong El Niño events of the 1980s and 1990s. The QIC record shows a small decline in accumulation during this period, further supporting this interpretation. However, without annually resolved dates, it is not possible to make any definite conclusions at this time.

Zone 1 at the top of core shows a shift to a more acidic, wetter conditions characteristic of a poor fen. It is difficult to make inferences about this shift because the QIC record ends in

2003, which is approximately where Zone 1 begins. However, one possibility is that this shift to wetter conditions may reflect a series of strong La Niña events that occurred from the late-1990s to 2010/2011. This interpretation is, again, based on the body of research that demonstrates the linkages between ENSO and precipitation in the tropical Andes. Clearly, though, more research is needed to study the relationship between bofedal diatom assemblages in the tropical Andes and ENSO.

Overall, diatoms preserved in peat in the Cordillera Vilcanota are most likely tracking moisture changes, highlighting their capacity to reconstruct long-term climate change in the Andes. This thesis provides insights into the potential diatoms may provide to study linkages between ENSO, SST, and regional climate in tropical Andes. Furthermore, <sup>14</sup>C dating of the bofedal demonstrates that peat growth in this region is very high and therefore demonstrates the importance of bofedales in carbon sequestration. If climate change continues unabated, it is possible that glaciers will disappear completely from the Cordillera Vilcanota and glacial-fed bofedales may be degraded, limiting their ability to act as biodiversity hotspots, carbon sinks, and natural water stores.

This thesis is the first to provide a detailed history of diatom assemblages preserved in peat in the Cordillera Vilcanota. It demonstrates that diatoms are indeed well preserved and likely responsive to moisture conditions that may reflect changes in glaciation, regional climate, and ENSO. Given the key ecosystem services bofedales provide, more research is needed to study the impacts of recent climate change on these ecosystems.

## Summary

- The overall objective of this thesis was to track the environmental history of a bofedal in the Cordillera Vilcanota, Peru, using diatoms and other siliceous indicators preserved in peat deposits.
- $^{14}\text{C}$  dating indicated that the peat core represents a history of the past 230 cal. yr. BP, indicating peat accumulation rates that are amongst the highest in the world. Similar rates of peat accumulation have been observed elsewhere in the tropical Andes.
- Diatom assemblages show that cool, alkaline, oligotrophic conditions were prevalent until the mid- to late-20<sup>th</sup> century, at which point the dominant diatom taxa indicate a shift to drier conditions. Following a period of aridity, the diatom assemblage shifts again, indicating wetter and more acidic conditions characteristic of a poor fen.
- The many diatom shifts during the past century may be a result of several factors, most notably glacial melt due to climate change and/or ENSO events. The QIC record also indicates that precipitation gradually increased over the 20<sup>th</sup> century, except for a small decrease in accumulation at the end of the record.
- These data demonstrate the sensitivity of diatom assemblages in peat bodies to changes in climatic variables and thus their potential to study long-term environmental change in the rapidly changing Andes.

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## Appendix

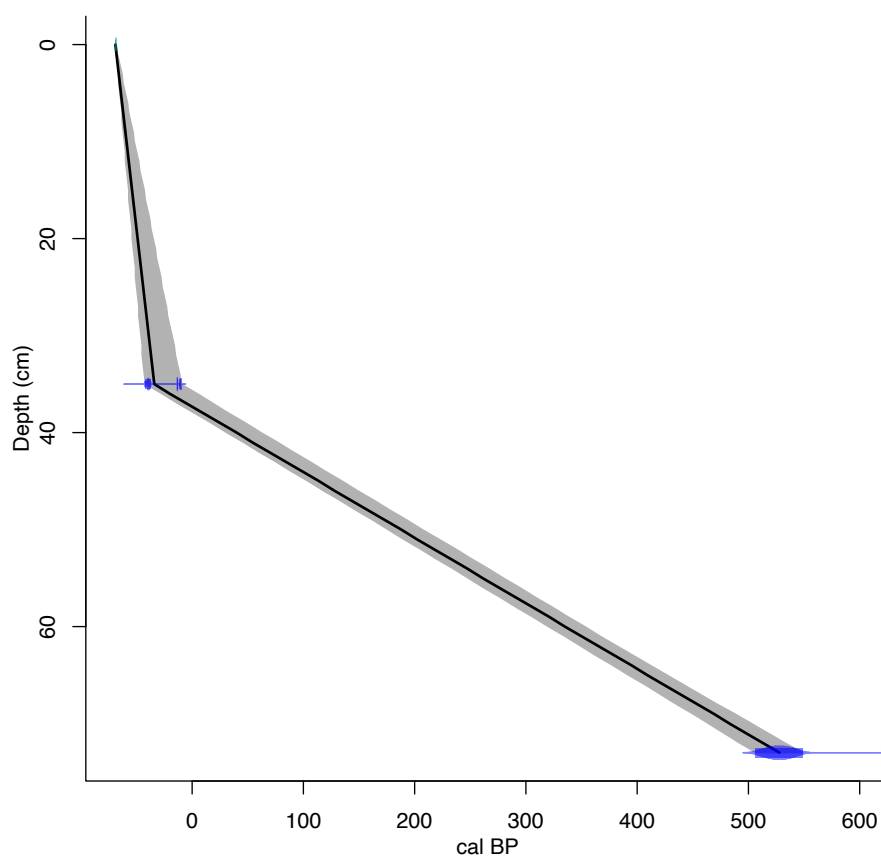


Figure S1: Clam  $^{14}\text{C}$  model for the lower bofedal core. The core shows a similar accumulation rate that is observed in the upper bofedal core.

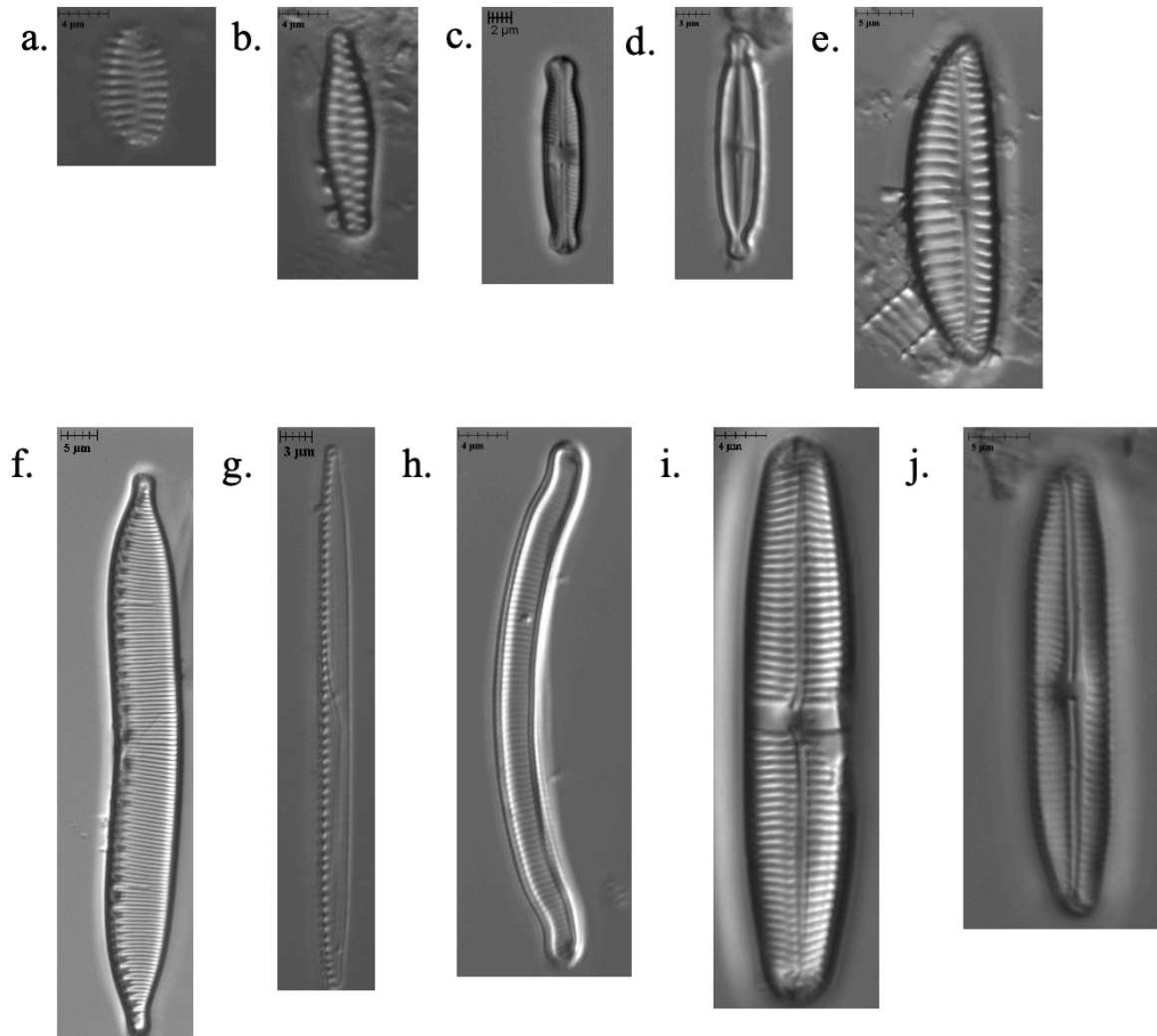


Figure S2: Common diatom taxa observed in the bofedal core. a) *S. pinnata*; b) *S. construens*; c) *S. cf. subnympharum*; d) *K. subtileissima*; e) *E. silesiacum*; f) *H. amphioxys*; g) *N. paleacea*; h) *E. arculus*; i) *P. cf. kuetzingii*; and j) *E. norvegicum*.