

METEORIC ^{10}Be IN LAKE SEDIMENT CORES AS A MEASURE OF CLIMATIC
AND EROSIONAL CHANGE

by

Rachel Elizabeth Jensen

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DEDICATION

For my life partner, Stephan, and my parents for their unending love and support.

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ABSTRACT

Developing tools that trace Earth-surface processes is necessary to quantify the complex controls on geomorphological, geochemical, and climate records. This thesis explores the potential of one such tool, meteoric ^{10}Be . The delivery of meteoric ^{10}Be to the surface varies with precipitation and its adsorption to sediment has proven useful in studies of erosion. These characteristics indicate that meteoric ^{10}Be in lake sediments varies under changing climate and changing sediment influx, making it a potential recorder of past climate and landscape processes. To examine the controls on meteoric ^{10}Be concentrations in lake sediments, we develop a model that predicts concentrations of ^{10}Be in lake sediments as a function of atmospheric flux, sedimentation rate, and terrigenous input. The model was applied to two published datasets of ^{10}Be profiles in lake sediments from different settings to assess the sensitivity of individual controls on ^{10}Be concentration. Results show that while a variety of environmental conditions influence ^{10}Be in lake sediments, these can be quantified with surprisingly simple parameters. Assessment of the relative importance of model parameters requires the comparison of ^{10}Be concentrations in well-dated lake cores to independent paleoenvironmental proxies. We further validate this model and explore the application of meteoric ^{10}Be in lake archives, by exploring a new system, Blacktail Pond in Northern Yellowstone, for which a wealth of paleoenvironmental data exists. We present new meteoric ^{10}Be data in the core, and compare to model predictions based on sedimentation rates of both autochthonous and allochthonous sediments and changes in the flux of meteoric ^{10}Be with precipitation. Surprisingly, patterns of measured ^{10}Be concentrations in Blacktail Pond sediments show little relationship to predicted concentrations, despite being of similar magnitude. Based on this analysis, we suggest that small lake systems may be most problematic for ^{10}Be analyses as they are most sensitive to changing ^{10}Be concentrations relative to changes in model parameters. This work provides a new quantitative framework to assess the control of sedimentation rates, inputs of allochthonous sediments, and hydroclimate in determining the ^{10}Be concentrations measured in lake sediments, and highlights the potential and limitations of meteoric ^{10}Be in quantifying past environmental changes.

CHAPTER 1

INTRODUCTION

Earth's surface, the facet of the planet with which we most directly interact, is shaped by the interactions between rock, soil, air, water, and life. Quantification of changes in these components and the interactions between them, especially over long timescales, is critical for understanding the evolution of this dynamic surface. Such an understanding is dependent on reconstructing long-term changes in the climate and the landscape. Developing tools that quantify characteristics of Earth-surface processes is critical for enabling understanding of how these processes may change over time.

One such tool, meteoric beryllium-10 (^{10}Be), shows promise as a proxy for paleoenvironmental change when measured in lake sediments. The delivery of this cosmogenic nuclide to the surface has been shown to increase with increasing precipitation (Figure 2.1) (e.g., Graly et al. 2011; Graham et al., 2003). In addition, the nuclide is strongly adsorbed to fine soils and sediments upon delivery to the surface (Graly et al., 2010). Thus, ^{10}Be concentrations in well-dated accumulated sediments may record past variations in precipitation and erosion within a catchment, and thus serve as a novel paleoenvironmental proxy.

Meteoric ^{10}Be is produced by spallation reactions in the upper atmosphere when cosmic rays produced by the shockwaves of supernovae collide with the nuclei of target atoms to produce several lighter particles, including ^{10}Be (Figure 1.1) (Lal and Peters, 1967; Masarik and Beer, 1999; Masarik and Beer, 2009). Meteoric ^{10}Be differs from the

production of in-situ produced ^{10}Be , which is created at known rates on Earth's surface within the crystal lattices of mineral grains (Figure 1.1) (Lal and Peters, 1967). Meteoric ^{10}Be is scavenged by precipitation and well mixed in the atmosphere (Heikkila et al., 2009) before delivery to the surface by rain, snow, or dry deposition (Figure 2.1) (McHargue and Damon, 1991). See Appendix A for a complete literature review summarizing our understanding of meteoric ^{10}Be .

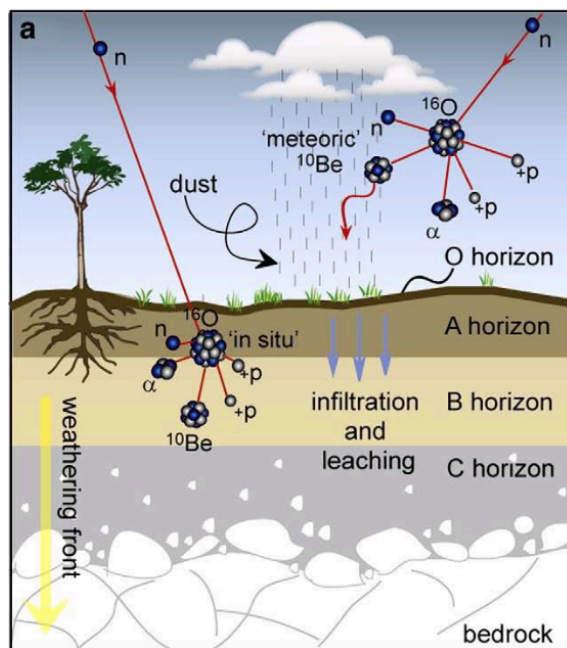


Figure 1.1. Production and delivery mechanisms of ^{10}Be and its incorporation into the soil (Figure from Willenbring and von Blanckenburg, 2010).

Generally, ^{10}Be is used in studies of weathering, erosion, and soil processes. Because in-situ ^{10}Be can quantify many of the same geomorphological characteristics as meteoric ^{10}Be , and because it has more easily quantified production rates, in-situ ^{10}Be is

used more widely (e.g., Dixon and Riebe, 2014). However, as progress is made in understanding the controls on the production of ^{10}Be in the atmosphere and its delivery to the surface, the potential of meteoric ^{10}Be as a versatile geochemical tracer in a wide variety of environments is becoming evident, though it is still underutilized.

Accumulated lake sediments are often used to help reconstruct terrestrial paleoenvironmental and paleoclimatic conditions, though they remain surprisingly underutilized sources of information regarding past geomorphic conditions. Sediment cores preserve evidence of geomorphic change, such as changes in the source or timing of dust delivery or large erosion events (Cohen, 2003). Several studies have measured meteoric ^{10}Be in lake sediments, showing the promise of this nuclide as an indicator of past environmental change (Horiuchi et al., 1999; Horiuchi et al. 2001; Belmaker et al., 2008; Belmaker et al., 2014; Berggren et al., 2010; Nilsson et al., 2011; Mann et al., 2012). Despite this promise, widespread use of the nuclide and robustness of insights are limited by the lack of a quantitative framework through which to interpret results. To address this need and identify and assess the relative importance of individual controls on the concentrations of ^{10}Be in lacustrine sediments, I present a simple, quantitative model that predicts ^{10}Be concentrations with depth in a sediment core based on lake sedimentation rate and the contributions of meteoric ^{10}Be from the atmosphere and from the catchment in both water and terrigenous material.

In Chapter 2, I present the model application to two published datasets that report ^{10}Be concentrations in dated lake sediments. These systems, Lake Baikal, Russia (Horiuchi et al., 1999) and former Lake Lisan, the Pliocene extent of the Dead Sea

(Belmaker et al., 2008), provide two distinct settings subject to different climates and landscape processes ideal for applying my model. Our application to these two datasets shows how ^{10}Be concentrations in lake cores may vary over long timescales both due to and despite changes in climatic and environmental forcings. Of the major controls on ^{10}Be concentrations, sedimentation rate governs the pattern of ^{10}Be distribution in a core, while the total magnitude of ^{10}Be concentration is a combination of the contribution of ^{10}Be to a core by atmospheric flux and terrigenous material or catchment runoff. However, changes in precipitation, which govern atmospheric delivery of ^{10}Be to a lake, have a limited impact on concentrations. Conversely, the input of ^{10}Be on terrestrial sediment may change concentrations dramatically, indicating that meteoric ^{10}Be might be a powerful proxy for paleoerosion and dust deposition. This is the first work to provide a quantitative framework for predicting ^{10}Be concentrations in accumulated sediments and constrains a multifaceted set of controls to a few simple parameters.

In Chapter 3, I present new data on meteoric ^{10}Be concentrations in the sediments of Blacktail Pond in northern Yellowstone. In conjunction with previous analysis of paleoenvironmental proxies, the ^{10}Be concentrations of Blacktail Pond offered the opportunity to test my model framework in a completely new system. This is one of only a few studies to compare measured ^{10}Be concentrations to a variety of paleoclimatic data. Published paleoenvironmental studies of the region and of the sediments of Blacktail Pond itself (see Appendix B) (Krause and Whitlock, 2013; Whitlock et al., 2012; Whitlock and Bartlein, 1993; Millspaugh et al., 2004) provide variables that fit the model parameters and, therefore, enabled modeling of ^{10}Be concentrations in Blacktail Pond

sediments before they were measured. Comparing modeled and measured ^{10}Be concentrations in this system provides insight into how and why measured ^{10}Be might vary from predictions based on knowledge of paleoenvironmental conditions.

The potential of meteoric ^{10}Be in paleoenvironmental records has been largely unexplored. This is among the first work to identify and quantify the primary parameters that contribute to ^{10}Be concentrations in accumulated sediments in vastly different settings undergoing large environmental changes. Additionally, measuring ^{10}Be in Blacktail Pond against well-dated, thoroughly studied paleoenvironmental records allows for comprehensive testing of this quantitative model framework in a new system and the viability of meteoric ^{10}Be as a paleoenvironmental proxy in light of documented environmental changes. As more information is uncovered concerning meteoric ^{10}Be , an increasing number of studies aim to utilize it as a tool for quantifying geomorphic change in a variety of landscapes and conditions. This work advances the analytical and quantitative use of this versatile and important nuclide in understanding Earth-surface processes.

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CHAPTER 2

A QUANTITATIVE FRAMEWORK FOR UNDERSTANDING PAST
ENVIRONMENTAL CHANGE USING ^{10}Be IN LAKE SEDIMENTSContribution of Author and Co-Author

Author: Rachel E. Jensen

Contributions: Designed quantitative model framework, found existing ^{10}Be datasets, applied model to datasets, analyzed data, and wrote the manuscript

Co-Author: Jean L. Dixon

Contributions: Designed conceptual model framework, helped design quantitative model framework, edited model applications, discussed results and implications, and edited the manuscript.

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Abstract

Meteoric ^{10}Be may offer unique insights into paleoenvironmental records and to long-term variations of erosion and climate. The atmospheric delivery of ^{10}Be to Earth's surface varies with precipitation, and its strong adsorption to sediment has already proven useful in studies of erosion. While several studies have used ^{10}Be concentrations in lake cores to understand past environmental change, robust insights are limited by the lack of a quantitative framework to assess controls on nuclide abundance in lake cores. We develop a simple model to predict the concentration of ^{10}Be in lake sediments with depth as a function of atmospheric flux, sedimentation rate, and terrigenous input. We apply our model to two published datasets of lake sediment ^{10}Be profiles in very different settings, Lake Baikal, Russia, and former Lake Lisan, Israel. We assess the sensitivity of individual model controls on ^{10}Be concentrations in these systems. In our quantitative reanalysis of both data sets, sedimentation rate is the primary control on the pattern of changing ^{10}Be concentration in sediment archives, due to the dilution of ^{10}Be by in-situ produced lake sediments. However, atmospheric flux and sedimentation rates alone are likely insufficient to explain sediment ^{10}Be concentrations. Furthermore, precipitation-driven changes in atmospheric flux have very little influence on ^{10}Be concentrations. In both systems, ^{10}Be must be additionally delivered to the lake from by surface transport from upland sources or by wind-delivered dust to account for all of the measured ^{10}Be , indicating that ^{10}Be may be a useful erosion proxy. We show that while a variety of environmental processes likely influence ^{10}Be concentrations in lake sediments, simple parameters can be used to model ^{10}Be concentrations and nuclide profiles may vary

widely solely due to intrinsic variables (e.g., lake sedimentation), with surprisingly little external controls on the system (e.g., changes in paleoerosion). This work provides a new framework to identify the control of sedimentation rates, sediment inputs, and hydroclimate in determining the ^{10}Be concentrations measured in lake sediments and to assess how and why they might deviate from model predictions. It highlights the potential for meteoric ^{10}Be as a useful tool for quantifying past environmental changes.

Introduction

Measurements of meteoric ^{10}Be in soil and water provide unique insights into diverse processes across critical zone interfaces. The production and delivery of meteoric ^{10}Be to the surface reflect interactions between air, water and soil, whereas its behavior in soil can help to quantify erosion and weathering rates, processes that often involve the interactions of soil, rock, and life. Two varieties of ^{10}Be are produced by different mechanisms: meteoric ^{10}Be is produced in the upper atmosphere via spallation by cosmic rays, while the production of in-situ ^{10}Be occurs in the lattices of mineral grains (Lal and Peters, 1967). Both meteoric and in-situ produced ^{10}Be can be used to quantify a wide range of critical zone processes. They accumulate on Earth's surfaces over time such that their concentrations reflect the amount of time a surface has been exposed to cosmic radiation (Willenbring and von Blanckenburg, 2010). Since ^{10}Be has a long half-life of 1.39 million years (Chmeleff et al. 2010), it has a long residence on the Earth's surface, and its concentrations in natural materials can be used to calculate a soil or surface ages, rates of erosion and weathering, and soil movement across a landscape. Meteoric ^{10}Be is of special interest to due to its delivery to the surface in precipitation and its adherence to

fine sediments. These characteristics indicate that concentrations of meteoric ^{10}Be in accumulated sediments may provide a record of environmental change.

The Delivery of Meteoric ^{10}Be to the Surface

Meteoritic ^{10}Be is produced in the upper atmosphere, after which it is deposited via both dry and wet processes (Field et al., 2006). In most mid- and low-latitude locations, delivery rate of meteoric ^{10}Be seems to vary strongly with precipitation (Figure 2.1; e.g., Graly et al., 2011, Graham et al., 2003). At lower latitudes, ^{10}Be delivery may reflect changes in the fraction of wet to dry deposition of ^{10}Be due to the scavenging of ^{10}Be by precipitation, possibly reflecting local climatic conditions (Heikkilä et al., 2013). Models that map global ^{10}Be deposition also show that deposition is high in regions where precipitation is high. Although few data correlates precipitation with ^{10}Be concentrations in terrestrial systems, past work has shown that meteoric ^{10}Be deposition can be predicted by latitude and average annual precipitation (Graly et al. 2011; Graham et al. 2003).

It is important to note a lack of consensus concerning the relationship of ^{10}Be to precipitation. Some settings show that periods of less precipitation result in higher concentrations of ^{10}Be , implying flux to the surface remains the same (Yiou et al., 1997). There is even evidence suggesting that ^{10}Be delivery in some environments is diluted by precipitation (Figure 2.1B), especially in settings with high precipitation (more than 6 cm/month), in which all the available ^{10}Be is scavenged from the atmosphere (Monaghan et al., 1985; Field et al., 2006; Willenbring and von Blankenburg, 2010). When considering ^{10}Be in sediments, however, concentrations vary over a relatively small range compared with the range of precipitation values; thus resulting ^{10}Be concentrations in

atmospheric precipitation are largely controlled by rainfall rate regardless of any dilution effect. If an average or constant ^{10}Be deposition rate is accounted for, precipitation should deliver more ^{10}Be to a given location until all of the ^{10}Be is scavenged from the atmosphere, at which point, ^{10}Be concentrations in accumulated sediments would plateau.

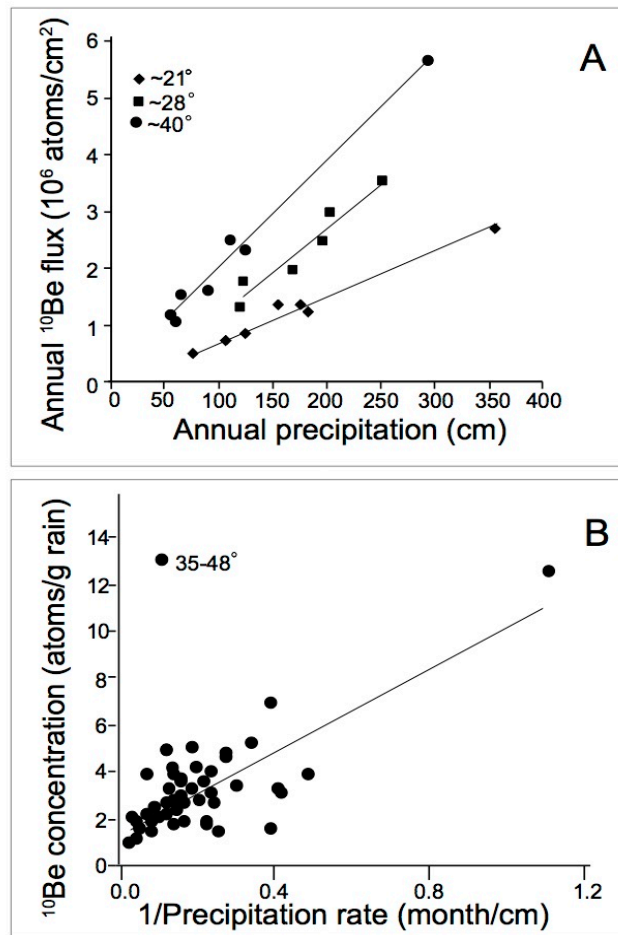


Figure 2.1. (A) Adapted from Graly et al. (2011). The atmospheric flux of ^{10}Be (10^6 A/ cm^2) generally increases with annual precipitation at varying latitudes. (B) In some landscapes, the concentration of ^{10}Be in rain (A/g rain) decreases with the amount of rainfall (data from Graham et al., 2004), however, because concentrations vary across smaller magnitudes than rainfall, this ‘dilution effect’ has limited impact on ^{10}Be fluxes.

Once delivered to the land surface as particle reactive $^{10}\text{BeO}_2$ or $^{10}\text{Be}(\text{OH})_2$, ^{10}Be binds readily to fine soil and sediment (Graly et al., 2010; Pavich et al., 1984). Meteoric ^{10}Be is retained on soil particles or incorporated as a cation into clays, organic matter, or oxyhydroxides under most natural chemical and mineralogical boundary conditions (Barg et al., 1997; Willenbring and von Blanckenburg, 2010). Dissolved ^{10}Be will only be lost to groundwater if the soils are exposed to sufficient water for mobilization. Thus, meteoric ^{10}Be is largely retained at the soil surface (Barg et al., 1997; Shen et al., 2004).

Meteoric ^{10}Be in Accumulated Sediments

Meteoric ^{10}Be has been widely used to understand a range of geomorphic processes, from the soil scale to the landscape scale. Concentrations of meteoric ^{10}Be in accumulated materials have most often been used to reconstruct temporal variations in atmospheric ^{10}Be production from high-latitude ice cores (e.g., Finkel and Nishiizumi, 1997; Muscheler et al., 2004; Vonmoos et al., 2006) and sediment cores (e.g., Berggren et al., 2010). Interpretation of ^{10}Be concentrations can be difficult in these records since the production signal is sometimes masked by climate influence on delivery (Wagner et al., 2000 and 2001).

The controls on ^{10}Be in accumulated sediments remain poorly quantified. Isotope concentrations in sediment may reflect both terrestrial transport and soil processes in addition to processes that control lake sedimentation and sediment chemistry, and atmospheric production of ^{10}Be . Therefore, it is not surprising that interpretations of ^{10}Be concentrations in sediment archives vary widely. Thus far, a quantitative framework has

not been developed to separate the mechanisms that govern meteoric ^{10}Be variability in sediment cores.

Only a limited number of studies have even explored the use of meteoric ^{10}Be in lake archives (Horiuchi et al., 1999; Horiuchi et al. 2001; Belmaker et al., 2008; Belmaker et al., 2014; Berggren et al., 2010; Nilsson et al., 2011; Mann et al., 2012). The majority of such studies attribute concentrations to fluctuations in production rates (Berggren et al., 2010; Nilsson et al., 2011; Mann et al. 2012), although they also note that the production signal may be complicated by hydrologic and climatic conditions that influence delivery of meteoric ^{10}Be to the lake. Raisbeck et al. (1981a) noted that ^{10}Be concentrations measured in lake sediments exceeded likely atmospheric delivery, and the additional ^{10}Be must come from eroded sediment.

We develop a simple framework to evaluate primary controls on meteoric ^{10}Be in accumulated lake sediments. This new model predicts the concentration of ^{10}Be as a function of atmospheric flux, sedimentation rate, and terrestrially delivered ^{10}Be (i.e. upland erosion or wind-blown dust). We apply the model to previously measured meteoric ^{10}Be profiles in lake systems from two distinct climate environments and latitudes. This quantitative framework constrains the complex processes and interactions to a few simple parameters so that future work aimed at reconstructing past environmental conditions can disentangle ^{10}Be signals resulting from environmental change from predicted concentrations based on our model parameters.

Approach: Modeling controls on meteoric ^{10}Be in lake systems

Model Framework

The concentration of ^{10}Be in sediment reflects the total flux of ^{10}Be delivered to the lake and the sedimentation rate, such that:

$$^{10}\text{Be} = \left(\frac{Q_{^{10}\text{Be}T}}{S} \right) \quad (\text{Equation 1})$$

Here, ' ^{10}Be ' is the concentration of ^{10}Be in sediment (atoms/g), ' $Q_{^{10}\text{Be}T}$ ' (atoms/cm²/yr) is the total flux of meteoric ^{10}Be to the lake system, and ' S ' (g/cm²/yr) is the total sediment accretion rate for the sediments in a core (cm/yr) multiplied by sediment density (g/cm³). In this formulation, lake sedimentation acts to dilute ^{10}Be concentrations delivered to the lake system. We assume decay is negligible, though a decay term could easily be added to the model if sediment records are sufficiently old ($\sim <50$ ka).

We identify several primary mechanisms that deliver ^{10}Be to the lake system, contributing to the total flux ($Q_{^{10}\text{Be}T}$) (Figure 2.2). Firstly, ^{10}Be is produced in the atmosphere, and directly delivered to the lake as wet or dry fallout. Secondly, atmospheric flux delivers ^{10}Be to the upland environment surrounding the lake, and so ^{10}Be may be delivered to the lake, as a function of hydrology, via runoff or groundwater inflow to the lake. Since ^{10}Be may tightly bind to upland soils and sediments, this hydrologic input may be limited to the high soil to water partition coefficient ($K=10^5$) (You et al. 1989; Pavich et al. 1986). Lastly, due to its reactive nature, it also enters the lake when terrestrial sediment with adhered meteoric ^{10}Be (inherited ^{10}Be) is carried to the lake by surface processes or wind.

The total meteoric ^{10}Be flux to the lake can be expressed as:

$$Q_{^{10}\text{Be}_T} = Q_{^{10}\text{Be}_A} + Q_{^{10}\text{Be}_H} + Q_{^{10}\text{Be}_E} \quad (\text{Equation 2})$$

Here, the total flux ($Q_{^{10}\text{Be}_T}$) is equal to the combination of atmospheric flux of ^{10}Be directly to the lake ($Q_{^{10}\text{Be}_A}$; atoms/cm²/yr), the hydrologic flux of ^{10}Be in water from the catchment ($Q_{^{10}\text{Be}_H}$), and the flux of inherited ^{10}Be arriving in the lake via terrigenous sediments ($Q_{^{10}\text{Be}_E}$).

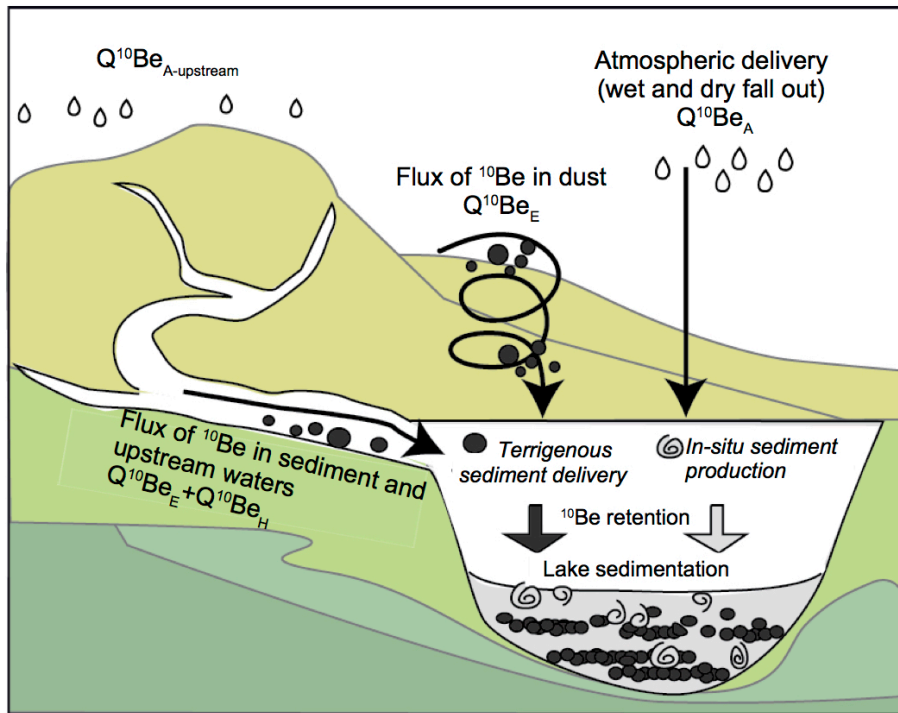


Figure 2.2. Conceptual model for the delivery of meteoric ^{10}Be to lake sediments. Meteoric ^{10}Be enters the lake from atmospheric fall out ($Q_{^{10}\text{Be}_A}$) and upstream delivery ($Q_{^{10}\text{Be}_A\text{-upstream}}$). ^{10}Be that is not retained in soils upstream may then be delivered hydrologically to the lake ($Q_{^{10}\text{Be}_H}$). Since ^{10}Be may also tightly bind to upland soils and sediments, and enter the lake through erosion or wind ($Q_{^{10}\text{Be}_E}$).

Atmospheric flux varies as a function of atmospheric production, transport, and deposition, though measured data in rainfall indicates that fluxes largely scale with

precipitation rate (Graham et al., 2003; Graly et al., 2011; Figure 2.1). The hydrologic flux ($Q_{10\text{BeH}}$) is limited by the integrated atmospheric flux across the entire upslope catchment and is maximized when there is little retention of ^{10}Be in upland sediments. Therefore, the maximum hydrologic flux can be described as:

$$\text{Maximum } Q_{10\text{BeH}} = \int Q_{10\text{BeA}(\text{upstream})} \cdot \quad (\text{Equation 3})$$

If atmospheric flux is uniform across the catchment, then:

$$\text{Maximum } Q_{10\text{BeH}} = Q_{10\text{BeA}} \times \frac{A_C}{A_L} \quad (\text{Equation 4})$$

Here, A_C is the upslope contributing area to the lake, and A_L is the lake area. Where there is little to no upland contribution of water to the lake, $Q_{10\text{BeH}} = 0$. Furthermore, if the catchment is soil-mantled and soil pH is sufficient to retain ^{10}Be in soils, then the hydrologic flux can be assumed negligible such that Equation 1 and 2 reduce to

$$^{10}\text{Be} = \left(\frac{Q_{10\text{BeA}} + Q_{10\text{BeE}}}{S} \right). \quad (\text{Equation 5})$$

The flux of ^{10}Be associated with inherited concentrations on terrigenous sediment ($Q_{10\text{BeE}}$) may arrive in the lake via upland erosion (attached to sediment transported in upland streams) or by dust deposition. To account for this source of ^{10}Be to the lake, one must know the amount of allocthonous terrigenous sediment delivered to the lake and its inherited ^{10}Be concentration. Therefore, it is necessary to identify what fraction of the sedimentation reflects terrigenous inputs versus in-situ biogenic or chemical sediment production. Accounting for both the atmospherically and terrestrially delivered ^{10}Be to the lake system, the concentration of meteoric ^{10}Be at a given depth in the core is:

$$^{10}\text{Be} = \left(\frac{Q_{10\text{Be}}}{S} \right) + \left(^{10}\text{Be}_E \times \frac{E}{S} \right) \quad (\text{Equation 6})$$

Here 'E' is the flux of terrigenous sediment ($\text{g}/\text{cm}^2/\text{yr}$), such that E/S reflects the fraction of total sedimentation rate accounted for by terrigenous sediment input (dust and/or upland erosion products). ' $^{10}\text{Be}_E$ ' (A/g) reflects the inherited component, which is the concentration of ^{10}Be attached to terrigenous sediment..

In a system in steady state with regards to meteoric ^{10}Be , all of the ^{10}Be delivered to the land surface from the atmosphere is removed via either runoff or erosion and delivered to the lake. In this kind of system, $Q_{^{10}\text{Be}E} = Q_{^{10}\text{Be}H}$, such that either flux, when added to $Q_{^{10}\text{Be}A}$, would account for all the ^{10}Be in accumulated lake sediments.

Model Validation Using Measured ^{10}Be Concentrations in Two Lake Systems

We apply our quantitative framework to two existing data sets of meteoric ^{10}Be in accumulated lake sediments to explore the degree to which simplified model parameters can account for and provide insight into controls on measured meteoric ^{10}Be . We focus on data collected from Lake Baikal by Horiuchi et al. (1991) (Table 2.1) and further interpreted in Horiuchi et al. (2001), and data collected from former Lake Lisan in Israel by Belmaker et al. (2008) (Tables 2.2 and 2.3) and further interpreted in Belmaker et al. (2011; 2014).

Comparison of ^{10}Be concentrations to changes in sedimentation rate with depth, represents a critical component of our model, so we selected data sets that present ^{10}Be concentrations at reported depths in independently dated sediments (Tables 2.1-2.3). Additionally, the studies and published datasets provide information that helps constrain a number of other important variables contained in Equation 6, including the delivery of terrestrially sourced and atmospherically sourced meteoric ^{10}Be .

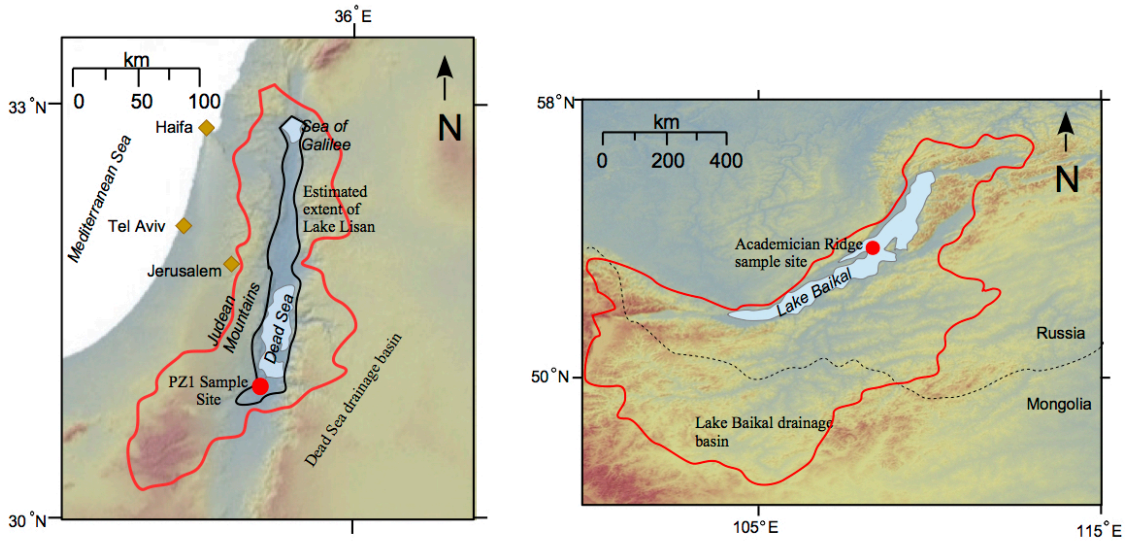


Figure 2.3. Lake Lisan and Lake Baikal drainage Basins. Lake Lisan, the Pleistocene extent of the modern Dead Sea, was located on the borders of modern Israel and Jordan at 31.56° N with an area of $2,500 \text{ km}^2$ and a catchment area of $40,000 \text{ km}^2$ (Belmaker et al., 2008). Lake Baikal lies in southern Siberia, at 53.42° N with an area of $\sim 31,700 \text{ km}^2$ and a catchment area of $\sim 560,000 \text{ km}^2$ (Demske et al., 2004). Extent of lake and catchment areas adapted from maps of Belmaker et al. (2014) and Gertcyk (2016).

The two study systems provide interesting contrasts in climate and landscape history, differences that aid the useful testing of our model. The sites lie at different latitudes and have distinct climatic regimes (Figure 2.3), thus experiencing different ^{10}Be production and delivery rates. They also reflect processes at different spatial scales; Lake Baikal lake and contributing catchment areas are an order of magnitude greater than those of former Lake Lisan (Figure 2.3). In the two sections below, we introduce each of the model systems and present insights gained from a comparison of modeled and measured ^{10}Be concentrations.

Model System 1: Lake Baikal

Background

As the oldest and deepest freshwater lake in the world, Lake Baikal provides a rich record of changing climate in Central Asia and has been the subject of abundant paleoclimate studies (e.g. Demske et al., 2004; Fagel and Boes, 2007; Goldberg et al., 2000; Goldberg et al., 2001a; Goldberg et al., 2001b; Goldberg, et al., 2010; Prokopenko et al., 2006; Tarasov et al., 2005). Horiuchi et al. (1999) measured concentrations of ^{10}Be with depth in a sediment core from the Academician Ridge in Lake Baikal (Table 2.1). The sediment ^{10}Be data, spanning a section of lake history over the last 23 ka, were compared with sediment texture (e.g., sand, silt, clay) and abundance of biogenic silica (Horiuchi et al., 1999) and later with geochemical data (Horiuchi et al., 2001) to characterize the environmental conditions that influence ^{10}Be distribution.

Application of the Model

Horiuchi et al., (1999) report ^{10}Be concentrations and ages for core segments that integrate across depth intervals. We calculated sedimentation rate (cm/yr) for the Lake Baikal core by dividing the average depth of each interval (Horiuchi et al. 1999) by the difference in age for each interval (Table 2.1). Note the general inverse correlation between sedimentation rate and ^{10}Be concentration (Figure 2.4).

We developed three sets of distinct model outputs based on assumptions on ^{10}Be fluxes to the lake. The first set of models assumes the only flux to the lake was via direct atmospheric input ($Q_{10\text{BeT}} = Q_{10\text{BeA}}$), such that there is no terrigenous or hydrologic input

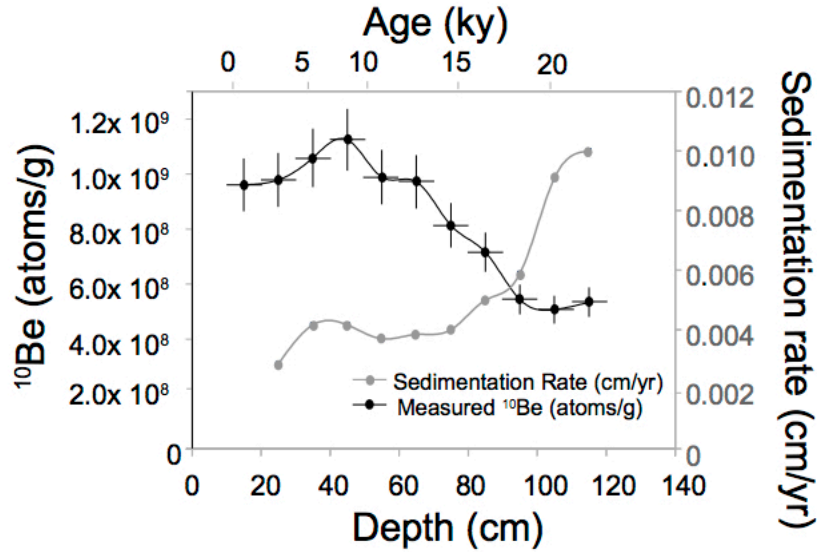


Figure 2.4. Sedimentation rate (cm/yr) plotted with the measured concentration of ^{10}Be in the Lake Baikal core (A/g). Concentration is represented in black. X-error bars on concentrations represent 10 cm sampling intervals and y-error bars represent a 10% error in value. The sedimentation rate is represented in grey on the right y-axis.

of ^{10}Be . We utilize the quantitative model presented by Graly et al. (2011), which calculates meteoric ^{10}Be flux as a function of latitude and annual precipitation, though any estimated flux for the region could be used as the flux term $Q_{^{10}\text{BeA}}$. We use several estimates of $Q_{^{10}\text{BeA}}$ to observe how sensitive the model is to atmospheric flux. We use the modern precipitation rate of 50 cm/yr to the Academician Ridge (Törnqvist et al. 2014), and bracket this value with two estimates of past precipitation (10% higher and 10% lower than modern values), to encompass estimates of the temporal variability of precipitation over the 20 ka sedimentary record. These result in a range of $Q_{^{10}\text{BeA}}$ between 1.02×10^6 and 1.13×10^6 atoms/cm²/yr for the low and high precipitation based estimates respectively. We also utilize an estimate of average global atmospheric flux of 1.21×10^6 atoms/cm²/yr for comparison (Monhagen et al. 1986; Masarik and Beer, 2009). Four

model curves show predicted patterns in ^{10}Be concentrations assuming only constant atmospheric inputs to the lake (Figure 2.5). It is important to note that we applied the average of bulk densities reported by Horiuchi et al. 1999 to estimate sedimentation rates in our model. Density assumptions have non-trivial impacts on modeled ^{10}Be concentrations due to the importance of sedimentation rates in model equations (see Supplementary Material 2B).

The second set of models account for ^{10}Be fluxes both directly to the lake and those integrating from across to the upstream area, such that $Q_{10\text{BeT}} = Q_{10\text{BeA}} + \int Q_{10\text{BeA}(\text{upstream})}$ (See Equations 1-3). We utilized the same range of precipitation-based fluxes and average global flux, and use 560,000 km² for the contributing area for Lake Baikal and 31,700 km² for the area of Lake Baikal (Demske et al., 2004). These modeled concentrations assume all atmospherically delivered ^{10}Be to the surrounding catchment is sourced to the lake system. Importantly, these models assuming upstream contribution are more than an order of magnitude higher than measured concentrations (Figure 2.5 inset).

The sets of modeled ^{10}Be concentrations thus far have assumed no terrigenous input, and are lower than ^{10}Be concentrations measured by Horiuchi et al. (1999). We develop a third set of models to account explicitly for input of inherited ^{10}Be with terrigenous sediments. To account for terrigenous input, we apply Equation 6, using the percentages of biogenic silica reported by Horiuchi et al. (1999) to estimate the percentage of sediment derived from upland settings (Equation 6). Previous research in the region suggests that biogenic silica in lake sediment is not significantly contaminated

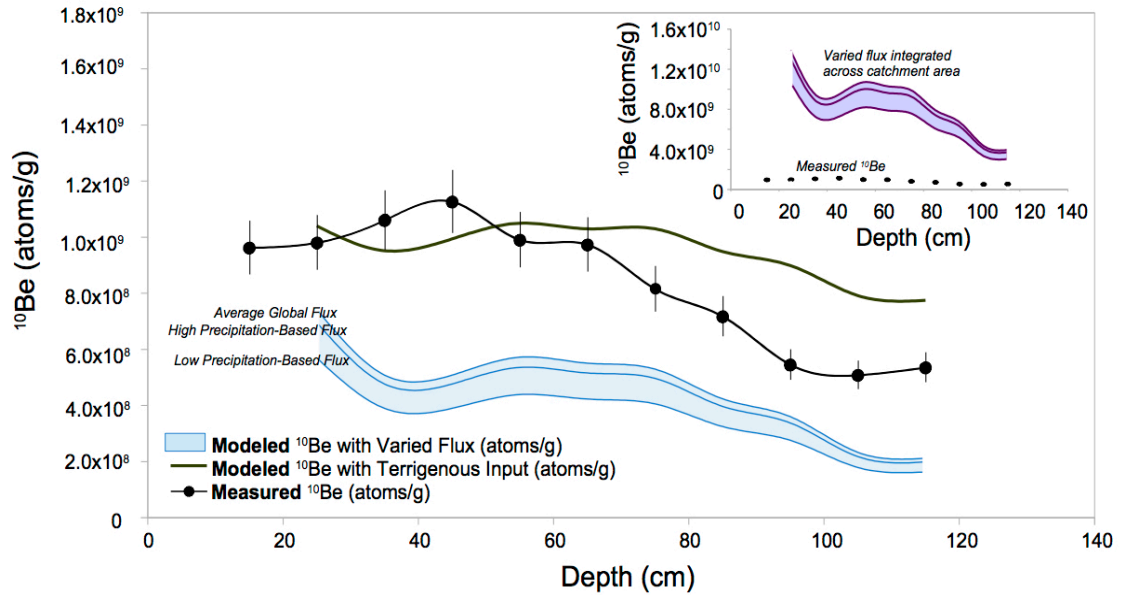


Figure 2.5. A comparison of our models to measured ^{10}Be concentrations in Lake Baikal. The concentrations of ^{10}Be in Lake Baikal measured by Horiuchi et al. (1999) are shown in black round symbols, as in Figure 4 (sample depth intervals not depicted here as x-error bars). First, we assume no terrigenous input (blue swath). The range concentrations of ^{10}Be predicted by our model, assuming no terrigenous input, utilize flux based on estimates of precipitation, as well as the global average ^{10}Be atmospheric flux. The models depicted in the inset utilize the same fluxes, but integrate them across the catchment area (purple swath), assuming all meteoric delivery upcatchment is delivered to the lake system. Next, we assume a terrestrial input of ^{10}Be derived from biogenic silica content and the concentration of ^{10}Be in incoming terrigenous sediment determined from core geochemistry (green solid line). We assume an average constant flux of ^{10}Be to the surface. Note that if the concentration of ^{10}Be in terrigenous sediment is varied by less than 1%, depending on the point, the differences between measured concentrations and modeled concentrations accounting for terrigenous sediment can be accounted for.

by clastic material (Goldberg et al., 1998). In the application of our models to the Lake Baikal data, we assumed that the weight percent of biogenic silica makes up the total of the autochthonous sediment in the core, such that E/S (Equation 6) is equal to the difference between the total sediment in the core and the percent biogenic silica (1-fraction of biogenic silica). These inferred E/S values ranged from 0.647 to 0.991 across the core. These calculated fractions of terrigenous material at each depth increment were

then multiplied by an assumed constant concentration of ^{10}Be on the terrigenous material. No independent measure of the inherited ^{10}Be concentration on incoming sediment was available for this system. However, adding a constant value of $^{10}\text{Be}_E = 4.8 \times 10^8$ atoms/g resulted in the closest fit (by reduction of sum of square residuals between measured and modeled concentrations). This best-fit concentration of ^{10}Be adhered to terrigenous material seems reasonable given meteoric ^{10}Be concentrations in soil tend to be of this order of magnitude (Willenbring and von Blankenburg, 2010; Graly et al. 2010) and similar concentrations have been measured at this latitude (Pavich et al. 1986). Importantly, all measured concentrations can be matched by adjusting $^{10}\text{Be}_E$ by less than 1%. Hence, differences between the modeled (with terrigenous input) and measured concentrations in Figure 2.5 can be explained by changing the amount of terrigenous material delivered to the lake over time (E/S), with only trivial changes in the inherited ^{10}Be concentration on this sediment ($^{10}\text{Be}_E$).

Discussion

Previous research by Horiuchi et al. (1999; 2000; 2001) attributed variations in ^{10}Be concentrations in a ~20 kyr-long record to climate-related changes in sedimentation rate, especially the dilution of the nuclide due to increased sediment flux to the lake during the arid glacial conditions.

In the application of our quantitative model to explore these data, we found that atmospheric flux to the lake alone accounts for an average of half of the measured ^{10}Be concentration in lake sediments (Figure 2.5). Therefore, an added component of flux is needed to achieve the ^{10}Be concentrations measured by Horiuchi et al., (1999), and we

model this added flux as one associated with terrigenous sediment inputs. Based on our models, terrigenous sediments and atmospheric fluxes likely contribute similar amounts of meteoric ^{10}Be to the total ^{10}Be inventory in the lake archives. We also find that the concentration of ^{10}Be in terrigenous material does not need to vary significantly (<1% of assumed average value) to fully explain changes in measured concentration of ^{10}Be in the core. Importantly, the principal control on ^{10}Be concentrations appears to be sedimentation rates, and we show that measured concentrations can largely be explained without invoking major changes in the sources of ^{10}Be to the lake system.

Model System 2: Lake Lisan

Background

Lake Lisan reflects the Pleistocene extent of the Dead Sea. Lisan sediments are widely thought to reflect regional hydroclimate (e.g., Bartov et al., 2002), and sediment cores from this region have been the subject of a variety of paleoclimatic, paleoseismic, and paleomagnetic studies (e.g. Marco et al., 1996; Marco et al., 1998; Machlus et al., 2000). The sediments consist of alternating laminae of aragonite and fine detrital sediments (Stein et al., 1997). Aragonite bands are thought to precipitate in response to bicarbonate supplied runoff, while bands of detrital sediment represent terrigenous dust from the Sahara and Negev Deserts (Stein et al., 1997; Frumkin and Stein, 2001). Belmaker et al. (2008; 2011; 2014) utilize these theories to understand the distribution of meteoric ^{10}Be over the past 60 kyr of accumulated sediments in the Dead Sea region. Belmaker et al. (2008) measured meteoric ^{10}Be concentrations within a sediment core of Lake Lisan (Tables 2.2 and 2.3).

Application of the Model

Belmaker et al. (2008) report ^{10}Be concentrations, as well as sediment age, from point samples in the core, rather than depth-integrated samples. They report ages for bulk sediment including both aragonite and detrital laminae (Figure 2.6B). The authors present core depth as heights relative to the bottom of the core, which we recalculated to reflect depth from the surface to be consistent with our prior analysis in the Lake Baikal system. We calculate sedimentation rate (cm/yr) by dividing the difference in depth by the difference in age for each point at which ^{10}Be concentration was reported. Belmaker et al. (2008) utilize a previously published age-depth model (Supplementary Material 2A). As in Lake Baikal, we observe concentrations of ^{10}Be measured by Belmaker et al. 2008 appear to follow an inverse pattern to that of sedimentation rates (Figure 2.6A).

As in Model System 1, we develop three sets of models for Lake Lisan based on data presented in Belmaker et al. (2008). The first set of models account only for the direct atmospheric flux of ^{10}Be to the lake system, such that $Q_{^{10}\text{BeT}} = Q_{^{10}\text{BeA}}$, assuming no terrigenous sourced ^{10}Be . We utilize several estimates of atmospheric fluxes (blue swath in Figure 2.7), including the modern ^{10}Be flux to the region of 7.5×10^5 atoms/cm/yr as reported by Belmaker et al. (2008) as well as the ^{10}Be flux predicted by latitude and precipitation (Graly et al., 2011). We calculate the precipitation and latitude-based flux for both the modern annual precipitation to the Dead Sea (10 cm/yr, resulting in a flux of 1.42×10^5 atoms/cm²/yr; Israel Meteorological Service) and to the Judean Mountains (50 cm/yr, resulting in a flux of 7.08×10^5 atoms/cm²/yr; Israel Meteorological Service), the major contributing area to the Dead Sea (Belmaker et al., 2008, 2011, and 2014) (Figure 2.7). These models assume constant atmospheric flux and a single sediment density

(Belmaker et al. 2008 assume bulk density averages 1.75 g/cm^2) to predict concentrations of ^{10}Be .

Concentrations of ^{10}Be in aragonite bands, reported by Belmaker et al. (2008), provide a useful test of assumed atmospheric flux values, assuming these concentrations reflect only atmospheric flux of ^{10}Be to the lake and the upland setting with no terrestrial input (Belmaker et al., 2008) (Figure 2.7). ^{10}Be concentrations predicted from our first model set – assuming solely atmospheric input – are consistently lower than measured concentrations measured in aragonite (Figure 2.7), accounting for only 15-50% of the total measured ^{10}Be (Figure 2.7). An average atmospheric flux of $1.7 \times 10^6 \text{ atoms/cm}^2/\text{yr}$ would be needed to match the average ^{10}Be in aragonite sediment (red curve in Figure 2.7). Note that the measured concentrations in aragonite bands tend to vary more strongly than both the measured concentrations in bulk sediments and our model outputs (Figure 2.7). This similarity between model and measured concentrations for bulk sediments may result from the fact that our models account for sedimentation rate of both detrital and aragonite sediments reported by Belmaker et al. (2008) and utilized the reported values for density from Belmaker et al. (2008) to calculate average bulk sediment.

Since models that incorporate only atmospheric flux directly to the lake are lower than measured concentrations, there must be additional ^{10}Be delivered to the lake, either hydrologically or as adhered to dust or riverine sediment. Our second set of models account for atmospheric ^{10}Be delivery directly to the lake, and also include the maximum estimated runoff contributed from the upslope catchment. Here, we incorporate both the

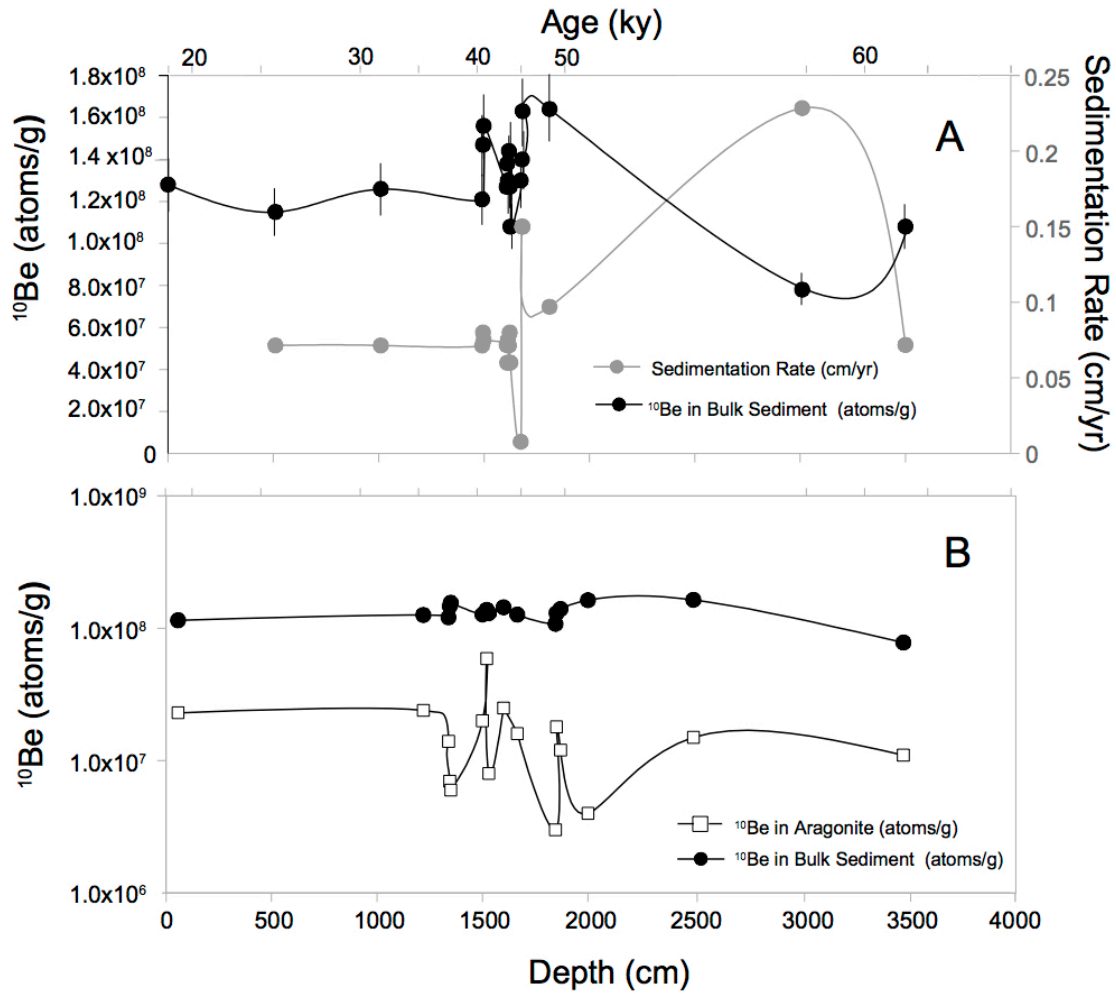


Figure 2.6. (A) Total measured ^{10}Be concentration (black) in Lake Lisan sediments. Sedimentation rate (grey) is displayed on the right y-axis. Y-error bars on bulk sediment ^{10}Be concentrations represent 10% error. (B) Measured ^{10}Be concentration in bulk sediment (black) and measured ^{10}Be concentration in aragonite laminae only (hollow squares) as reported in Belmaker et al. (2008). Log axes in panel B highlight concentration differences. 10% error bars are contained by symbols.

40,000 km^2 drainage basin and 2,500 km^2 lake surface area (Belmaker et al. 2008) to

account for maximum atmospheric flux of ^{10}Be to the upland catchment area (see

Equations 1-3). Recall, here $Q_{10\text{BeT}} = Q_{10\text{BeA}} + \int Q_{10\text{BeA}(\text{upstream})}$ (See Equations 1-3). We

note that the measured and modeled ^{10}Be concentrations in bulk sediments are roughly

matched, and occur at the same order of magnitude (Figure 2.7 inset).

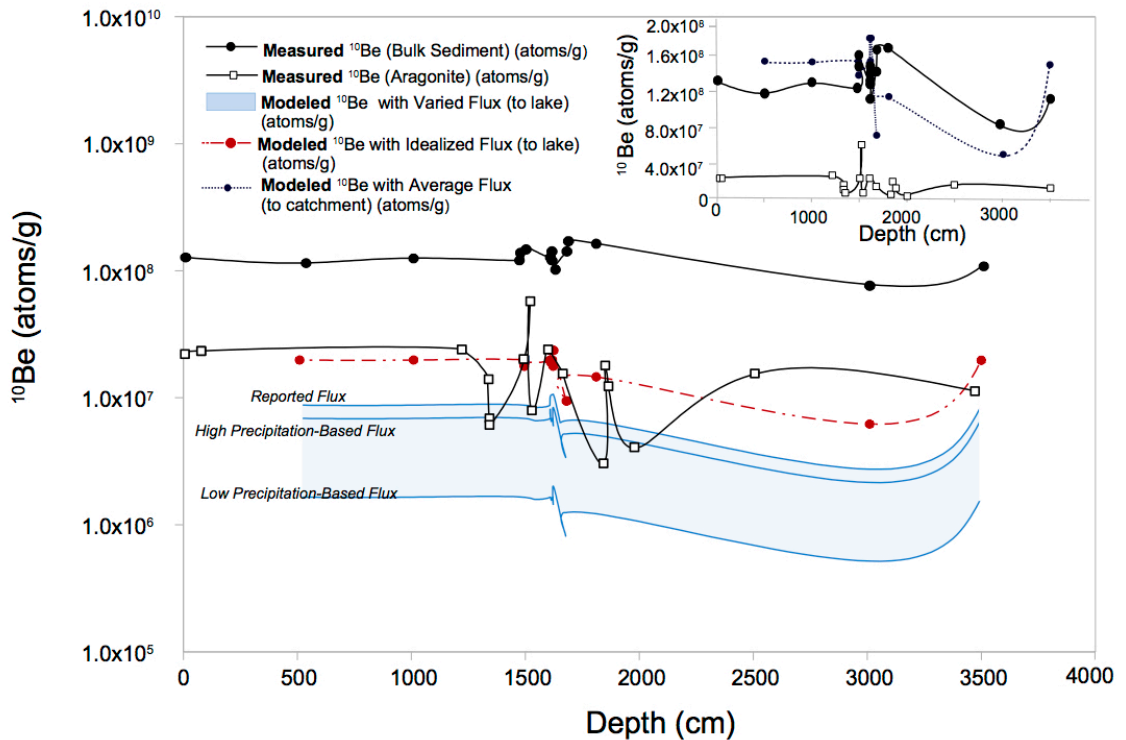


Figure 2.7. The measured concentrations of ^{10}Be in former Lake Lisan measured by Belmaker et al. (2008) (black circles represent bulk sediment concentrations and hollow squares represent aragonite concentrations) and the range concentrations of ^{10}Be predicted by our model, assuming no terrigenous input, assuming different estimates of flux (blue swath). The range concentrations of ^{10}Be predicted by our model utilize flux based on estimates of precipitation, as well as the atmospheric ^{10}Be flux reported by Belmaker et al. (2008). We include our idealized flux (red with round symbols), which accounts for the levels of ^{10}Be in aragonite-only sediments. We assume no terrigenous input for these models. 10% error is obscured by symbols. Inset: The model (dotted dark blue line with round symbols) depicted in the inset average these fluxes, but integrate them across the catchment area.

The third set of models account for both atmospheric inputs and the input of ^{10}Be associated with deposited terrigenous sediments (Figure 2.8). Belmaker et al. (2008; 2011; 2014) suggest that dust delivered to the lake from upland wadis during flooding

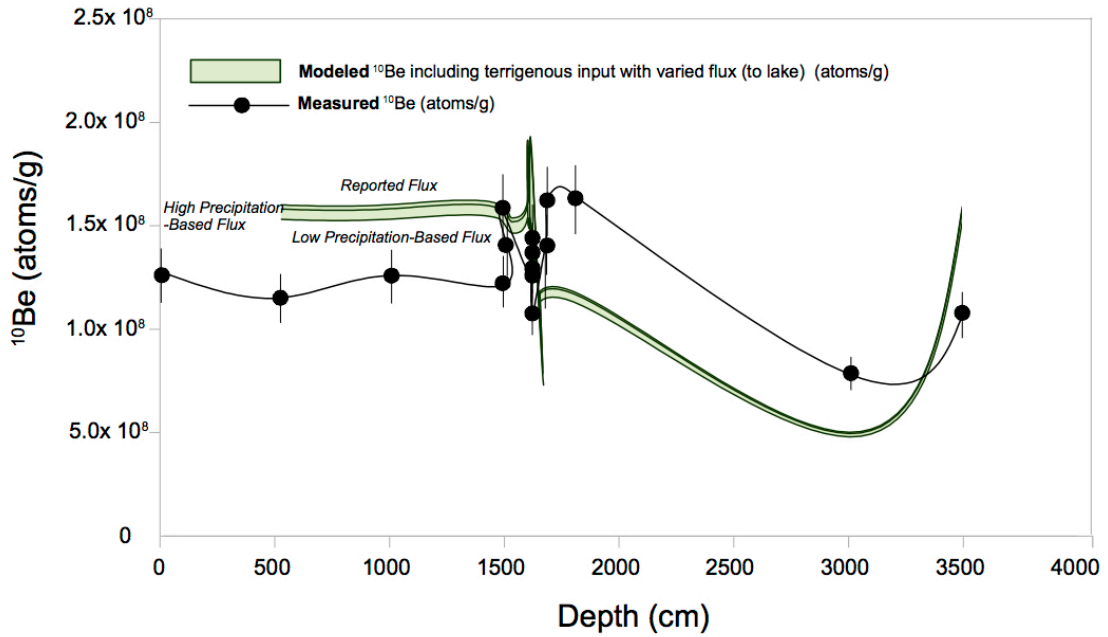


Figure 2.8. Measured concentrations (black circles) and modeled concentrations (green swath) of ^{10}Be . Modeled concentrations are based on three estimates of atmospherically delivered ^{10}Be to the surface (assumed to be constant) and a terrestrial input of ^{10}Be derived from measured dust fluxes and concentrations of ^{10}Be in dust.

contributes to the ^{10}Be concentrations in the lake. They report both an average dust flux and an average concentration of ^{10}Be in the dust of the region. To account for both terrestrial and lake-produced sediment contributions to total meteoric ^{10}Be concentration in accumulated lake sediments, we estimate the fraction of terrigenous sediment in the core, reported as E/S (where E is terrestrial input in $\text{g}/\text{cm}^2/\text{yr}$ and S is the product of sedimentation rate and density (see Equation 5). We used Belmaker's reported dust deposition rate of $0.11 \text{ g}/\text{cm}^2/\text{yr}$ to the region. Combined with their reported ^{10}Be concentration in dust of 1.18×10^8 atoms/g, we calculated the dust contribution of ^{10}Be to the lake (atoms/g lake sediment), assuming these parameters. These models allow changes in terrigenous sedimentation (E/S), but assume a constant inherited ^{10}Be concentration of these sediments ($^{10}\text{Be}_E$) and a constant atmospheric flux (using the same

three estimates in model set 1). Comparison of between measured concentrations and those modeled including terrigenous input are provided in Figure 2.8.

Discussion

After measuring ^{10}Be in various upland settings that likely contributed sediment to Lake Lisan, Belmaker et al. (2008; 2011; 2014) propose that unless there is a dramatic change in atmospheric ^{10}Be production rates, as during the Laschamp geomagnetic excursion around 40ka (~1500-2000 cm depth), variations of ^{10}Be in Lake Lisan sediment reflect hydrological conditions in the watershed and subsequent lake chemistry (Supplementary Material 2D) (Belmaker et al., 2014).

Application of our model framework to the Lake Lisan data provides important insights into controls on ^{10}Be concentrations in this system. Firstly, we find lake sedimentation has a dilution affect on ^{10}Be concentrations in the core, and exerts important controls on the temporal variability of ^{10}Be in sediments (Figure 2.6). Secondly, as found in the Lake Baikal study, atmospheric delivery of ^{10}Be directly to the lake (where $Q_{10\text{BeT}} = Q_{10\text{BeA}}$) is insufficient to account for the measured ^{10}Be in Lake Lisan sediments, even when we use high estimates of past precipitation (Figure 2.7). Indeed, precipitation has relatively little impact on concentrations of ^{10}Be (Figures 2.7 and 2.8), though it can account for as much as 20% variability in modeled concentrations (Note blue swath in Figure 2.7). However, as noted before, this set of models, regardless of $Q_{10\text{BeA}}$ flux assumed, produce ^{10}Be concentrations well below the magnitude of the measured concentrations, and atmospheric fluxes directly to the lake account for <20% (and generally much less than) of measured concentrations from Belmaker et al., (2008).

Lastly, modeled ^{10}Be concentrations in Lake Lisan must include ^{10}Be inputs in excess of atmospheric fluxes, potentially hydrologic or associated with incoming sediment. Lake-produced aragonite bands show only $\sim 20\%$ of the ^{10}Be found in bulk sediments (Figures 2.6 and 2.7) and are only slightly higher than modeled concentrations associated with atmospheric flux directly to the lake (model set 1). Considering these aragonite precipitates would account for all dissolved ^{10}Be , both delivered directly to the lake via precipitation *and* via hydrology, we can assume that this hydrologic flux must be quite small. Therefore, it is likely that the added flux comes in the form of ^{10}Be adhered to incoming sediments and dust, such that $Q_{^{10}\text{Be}_T} = Q_{^{10}\text{Be}_A} + Q_{^{10}\text{Be}_E}$ (Model set 3; Figure 2.8). Utilizing dust fluxes and associated inherited ^{10}Be concentrations ($^{10}\text{Be}_E$) reported by Belmaker et al. (2008; 2011; 2014), we note a compelling match between measured and modeled concentrations. We note, however, that ^{10}Be patterns in our models deviate from measured concentrations in two important ways. Modeled concentrations are higher than actual concentrations before 1500 cm depth and lower than measured concentrations after 2000 cm depth. This implies a change in environmental conditions and/or the flux of ^{10}Be during this time period that is not predicted directly from changes in sedimentation rate alone.

It is interesting to note that when contributing area is accounted for, modeled concentrations roughly match measured concentrations (Figure 2.7 inset). Most ^{10}Be atmospherically delivered to the surrounding catchment is retained in soils. If we were to consider an upland environment in steady state with respect to ^{10}Be , where atmospheric inputs were matched by erosive outputs, then the flux of $Q_{^{10}\text{Be}_E}$ to the lake in the form of

terrigenous sediments would match the formulation of $Q_{10\text{BeH}}$ expressed in Equations 3 and 4. Therefore, it appears that measured concentrations of ^{10}Be in Lake Lisan sediments are roughly consistent with a landscape in isotopic steady state.

Conclusions

We develop a set of equations to provide a quantitative framework through which to interpret measured concentrations of ^{10}Be in accumulated lake sediments. Applying our model to ^{10}Be data sets from lakes at distinct settings and latitudes, we identified the simple, predictable controls on the concentrations of ^{10}Be in accumulated lake sediments. Based on our results, ^{10}Be in accumulated sediments provides limited direct insight into the magnitude of precipitation changes, and hence, is likely only an indirect proxy for climate change. Instead, ^{10}Be concentrations reflect both in-lake sediment production and upland erosion, both of which may indirectly relate to changes in the hydroclimate system. Data from both Lake Lisan and Lake Baikal suggest that sedimentation rate is the primary control on depth patterns of ^{10}Be concentrations (Figures 2.4 and 2.6). Sedimentation rates themselves may represent a variety of hydrologic factors and terrestrial soil and dust transport, which in turn may also reflect local climate. Generally, ^{10}Be concentrations were diluted by in-lake sediment production. This relationship is expected unless there is a significant change in flux due to changes in ^{10}Be production (e.g., of the magnitude of the Laschamp event) or unless incoming terrigenous material is especially high in ^{10}Be ($^{10}\text{Be}_E$) and contributes a non-trivially to lake sedimentation or there is an event that contributes a large amount of terrestrial sediment (E/S) in a short period of time. Changes in these variables explain how modeled concentrations deviate

from measured concentrations (Figure 2.5 and 2.8), demonstrating the power and limitations of the model. Importantly, using this quantitative framework, the variations in ^{10}Be concentration caused by sedimentation can be accounted for, making it possible to test to what degree measured ^{10}Be concentrations may reflect other variables such as erosional inputs.

We find that atmospheric fluxes contribute a greater proportion of total ^{10}Be in the sediments in Lake Baikal than in the Lake Lisan system. Though the Judean Mountains of Israel and Lake Baikal receive the same annual precipitation, Lake Baikal is at a higher latitude. Atmospheric fluxes at high latitudes (such as in the region of Lake Baikal), would likely contribute a greater component of ^{10}Be to total measured concentrations than at low latitudes. However, in both the Lake Baikal and Lake Lisan systems, atmospheric fluxes directly to the lake are insufficient to explain measured ^{10}Be concentrations in sediments (Figures 2.5 and 2.7). The addition of dissolved ^{10}Be in runoff from the contributing area or ^{10}Be associated with particulate input from terrestrial sources must be considered, and model outputs show the degree to which these inputs can be assumed to be steady or temporally variable.

Several variables are critical for our model, and therefore should be included in future studies seeking to use meteoric ^{10}Be in sediments as records of past environmental change. First, sediments must be well dated, as calculated sedimentation rates have a significant influence on model concentrations. It is also necessary to have a good understanding of sediment composition and density (see further discussion in Supplementary Material 2B). Data that quantifies the terrigenous and in-situ-produced

components to lake sedimentation provides independent constraints on model parameters (E/S). Additionally, hydrology of the contributing area may play a role in accumulated ^{10}Be concentrations, but its influence will be highly variable depending on lake size, mixing, hydrologic connectivity and the precipitation gradient in the region. If at all possible, measuring ^{10}Be directly in in-situ produced sediments (e.g., aragonite) allows useful estimation of atmospheric fluxes directly.

Not all of these variables were available in the two data sets we selected, which represented the best available systems for our model validation. Nevertheless, comparison with these existing data showed our model can be used to constrain the degree to which measured data can be accounted for simple assumptions regarding ^{10}Be fluxes and inputs. Models can thus aid in quantitative analysis of the sensitivity of meteoric ^{10}Be to environmental changes including past hydroclimate, erosion regimes, and dust deposition. These complex factors can be quantified with a few simple parameters: sedimentation rate, density, and terrigenous sediment fraction. Ignoring such dominant parameters as sedimentation rate, especially without a quantitative framework, could lead to the misinterpretation of paleorecords. Future work should aim to continue to test this model in a wider variety of lake environments, latitudes and in areas where dramatic changes in precipitation may have influenced the atmospheric delivery of ^{10}Be , especially on long timescales.

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Supplementary Material

A Quantitative Framework for Understanding Past Environmental Change Using ^{10}Be in Lake Sediments, by Jensen and Dixon

Supplement 2A: Explanation of Sedimentation Rates Used in the Lake Lisan Studies

Supplement 2B: The Importance of Density in Our Proposed Model

Supplement 2C: Summary of Findings by Horiuchi et al. (1999, 2000, 2001)

Supplement 2D: Summary of Findings by Belmaker et al. (2008, 2011, 2014)

Supplement 2A: Explanation of Sedimentation
Rates Used in the Lake Lisan Studies

Based on our modeling, sedimentation rates exert a strong control on the pattern of meteoric ^{10}Be concentration in the accumulated lake sediment. Therefore, we analyzed the sensitivity of our model to different sediment rate models, specifically focusing on those rates calculated here, and the ones associated with the age-depth models first presented by Hasse-Schramm et al. (2004) and used by Belmaker et al. (2008).

Belmaker et al. (2008) report the ages of sediment at different heights in the sediment the researchers collected from the PZ-1 section of the Lisan Formation, just south of the modern Dead Sea. Height is obtained from measuring the distance from the base of the formation. Ages reported by Belmaker et al. (2008) are calculated from the age-height equations described by Hasse-Schramm et al. (2004). Hasse-Schramm et al. (2004) detail a U-series chronology of the PZ-1 section of the Lisan Formation. Using U-Th ages of aragonite in the formation, the study designates stratigraphic units by how they correspond to Marine Isotope Stages and derives four age-height models to describe different height intervals in the section.

We correlate the ages reported by Belmaker et al. (2008) to the Hasse-Schramm et al. (2004) regression equations to determine the sedimentation rate applicable to each of the measurement by Belmaker et al. (2008) and to utilize these rates in our model (See Equation 1). For each point reported by Belmaker et al. (2008) from 3505 cm to 1880cm height, Belmaker et al. (2008) use the age regressions for the upper section, resulting in a sedimentation rate of 0.0714 cm/yr. Then, between 1831 cm and 1822 cm the age regression for the lower section is used, resulting in a sedimentation rate of 0.1578 cm/yr.

After this, the alternative age regressions for the lower Middle member and Lower member of the Lisan formation are used (see Hasse-Schramm et al., 2004). This means from 1694 cm to 495 cm height, the sedimentation rate is 0.2288 cm/yr and for 5cm, the sedimentation rate is 0.0714 cm/yr again. For our model analysis, as presented above, we calculated sedimentation rate by dividing the change in depth by the change in age for each point measured by Belmaker et al. (2008). This does, however, determine the sedimentation rate between each measurement, rather than the slope of the line at that point.

Note that because of the nature of our model, we had to convert the heights reported by Belmaker et al. (2008) to depths by assigning the highest height of 3505 cm a depth of 0 cm and determining the depths of the lower heights by subtracting each from 3505. In the comparison of the concentrations of ^{10}Be stemming from Equation 1 with this calculated sedimentation rate and the concentrations of ^{10}Be stemming from Equation 1 with the sedimentation rates from each of the four age-height models used by Belmaker et al. (2008) to determine age, there are significant differences (Figure 2.9). At around 1500 cm depth, the modeled ^{10}Be concentrations with the calculated sedimentation rate make a sharp increase, while the ^{10}Be concentrations from the reported sedimentation rates remain the same until the next point, after which they decrease markedly at ~ 3500 cm depth to a consistent low rate. These concentrations are lower than those predicted by the model with the calculated sedimentation rates. Additionally, the model utilizing the sedimentation rates we calculated are closer in trend to the measured ^{10}Be concentrations (Figure 2.9).

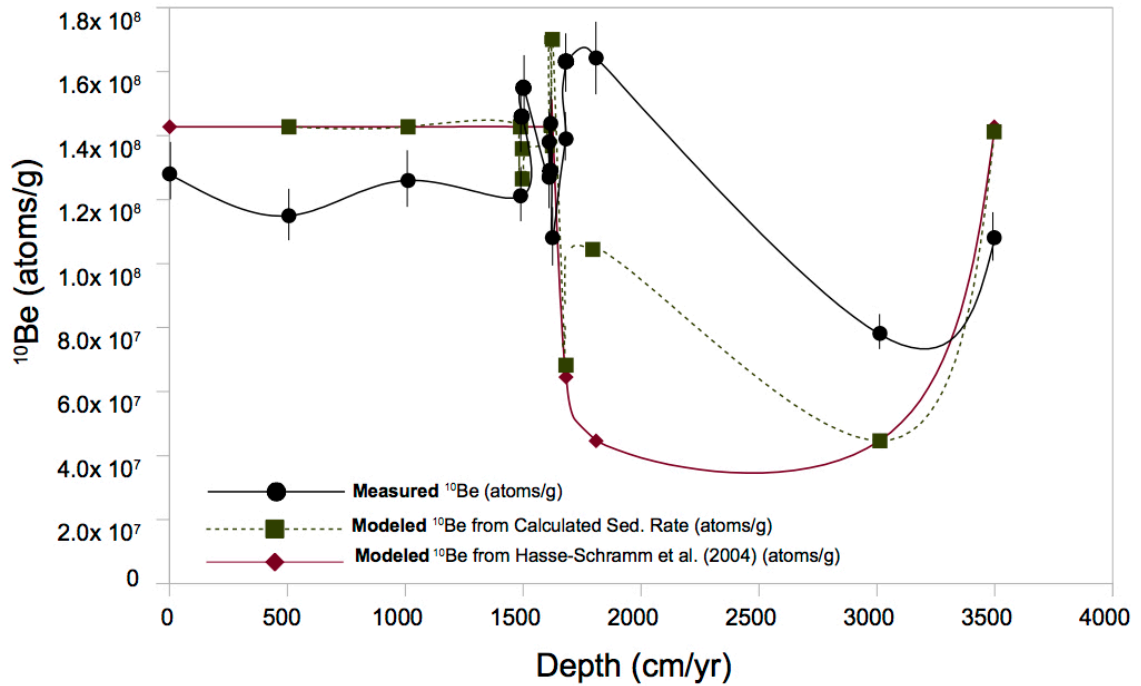


Figure 2.9. Comparison of Measured ^{10}Be concentration (black circles) to modeled concentrations using our calculated sedimentation rate (green with square symbols) and the step-wise sedimentation rate obtained from the age-height models reported by Hasse-Schramm, et al., 2004 and utilized by Belmaker, et al., 2008 to calculate age (red with diamond symbols). Y-error bars on measured concentrations represent 10% error.

We note several similarities in the two models, especially before 1500 cm. The model that utilizes sedimentation rates derived from the age-height models in Hasse-Schramm et al. (2004) assumes a constant sedimentation rate for long periods of time and assumes a step-wise change between each sedimentation rate. While the age-height regressions are adequate for the purposes of determining age for most studies, these assumptions may not allow for sensitive enough analysis to understand the variable environmental controls on sedimentation rate, and in this case, the controls on the patterns of ^{10}Be accumulation. So while we utilize the ages calculated by Belmaker et al. (2008), to calculate new sedimentation rates to use in our model, our calculations

encompass the changes in sedimentation rate between each point measured by Belmaker et al. (2008). As a result, the model utilizing the calculated sedimentation rate predicts a pattern of ^{10}Be concentration much closer to the actual ^{10}Be concentrations measures.

Ideally, if one were to use our model to predict ^{10}Be with depth in accumulated lake sediments, one would report the concentration of ^{10}Be and sedimentation rate for a given age-depth interval, rather than at a single point or between points.

Supplement 2B: The Importance of Density in Our Proposed Model

An important component of our model (Equations 1-3) is the term S ($\text{g}/\text{cm}^2/\text{yr}$). It is the product of sediment density (g/cm^3) and sedimentation rate (cm/yr). Because sediment density is often reported for large sections of core, assuming sudden, step-wise changes to account for different sediment types, we have generally elected to use total bulk density in our model. This way, the term is a constant. In Horiuchi et al. (1999), however, bulk density is measured for each age-depth interval reported. We modeled ^{10}Be concentrations using both the average of these densities ($0.571 \text{ g}/\text{cm}^3$) and with the changing densities reported for each interval (these range from 0.39 to $1.11 \text{ g}/\text{cm}^3$).

The concentrations of ^{10}Be modeled with changing density shows more similarity in trend to the measured ^{10}Be concentrations than the ^{10}Be concentrations modeled with average bulk density (Figure 2.10). They are even closer in magnitude to the measured concentrations. In this instance, using a constant average density results in an S value higher than the S derived from changing density from 5-70 cm depth. Between 70-110cm depth, S derived from constant average density is lower than the S derived from changing bulk density. Between 110 and 120 cm, S derived from constant density is higher again. This results in the differences in trends in the modeled concentrations in Figure 3.10. It is important to note that the densities reported by Horiuchi et al. (1999) generally increase with depth. This seems consistent with the observation that the percentage of biogenic silica decreases with depth. However, density over even a 10 cm span of core could vary significantly, especially in varved sediments. While sediment composition likely does exert control on ^{10}Be concentration, sedimentation rate and the percentage of terrigenously derived sediment seem to exert the most control.

We elected to use the ^{10}Be concentrations modeled from average bulk density to evaluate the effectiveness of our model. This accounts for the variability in sediment composition since it averages out density and allows for evaluation of other controlling factors. Additionally, because we calculated sedimentation rate between points, rather than for each interval (only the depth interval was reported in Horiuchi et al. (1999), not age interval), using average density, rather than the density of a single interval partially controls for this variability.

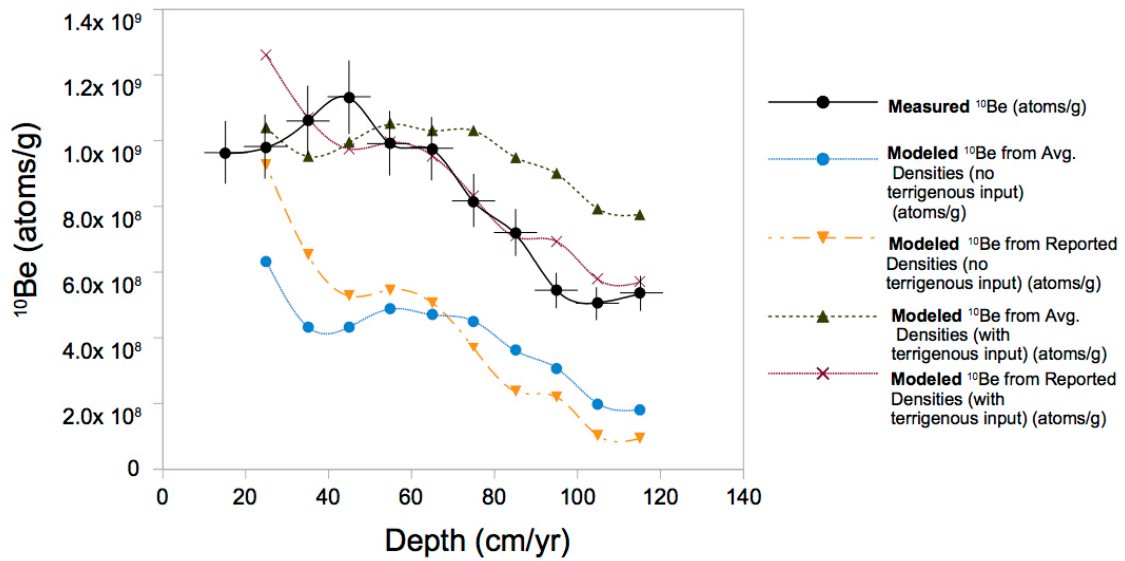


Figure 2.10. Measured concentrations of ^{10}Be (black circles) and modeled concentrations utilizing average bulk density, with no terrigenous input (blue with circle symbols) and with terrigenous input (green with triangle symbols) and density of sediment in each depth interval, without terrigenous input (yellow with triangle symbols) and with terrigenous input (red with x symbols). Y-error bars on measured concentrations represent 10% error and x-error bars represent 10 cm sampling intervals.

Supplement 2C: Summary of Findings by
Horiuchi et al. (1999, 2000, 2001)

As the oldest and deepest freshwater lake in the world, Lake Baikal provides a rich record of changing climate in Central Asia and has been the subject of abundant paleoclimate studies (e.g. Demske et al., 2004; Fagel and Boes, 2007; Goldberg et al., 2000; Goldberg et al., 2001a; Goldberg et al., 2001b; Goldberg et al., 2010; Prokopenko et al., 2006; Tarasov et al., 2005,).

A study by Horiuchi et al. (1999) examines the variation of ^{10}Be concentration with depth and sediment composition in a core from the Academician Ridge in Lake Baikal spanning the last 23,000 years (Table 2.1). Studies by Horiuchi et al., (2000; 2001) evaluate ^{10}Be concentration in a core from the same site spanning the last 150 ka. The authors compare ^{10}Be data with data describing sediment type – characterized as sand, silt, clay, and biogenic silica (Horiuchi et al. 1999) and later with 150 ka of geochemical data (Horiuchi et al., 2001) to characterize the environmental conditions that influence ^{10}Be distribution.

The researchers show that ^{10}Be concentrations are inversely related to sediment accumulation rate as well as clay content, but positively correlated with sand and biogenic silica content (Horiuchi et al., 1999). They calculate ^{10}Be flux *to the lake* ($\text{A}/\text{cm}^2/\text{yr}$) as a function of concentration, bulk density, and sedimentation rate and suggest that high ^{10}Be fluxes accompanied by low concentrations of ^{10}Be are indicative of large quantities of ^{10}Be poor sediment, primarily clay, being delivered to the lake during the last-glacial period likely as aeolian deposits. Conversely, higher concentrations of ^{10}Be at lower fluxes of ^{10}Be to the lake are associated with less dust delivery during the

wetter post-glacial stage. Later studies show that glacial periods show lower concentrations of ^{10}Be while inter-glacial periods show higher concentrations and support the idea that ^{10}Be concentrations are diluted by higher sediment input. ^{10}Be concentrations are compared with concentrations of a suite of elements, some of which are associated with warmer conditions and high biogenic silica levels (the “warm stack”) and some of which are associated with cold conditions and low biogenic silica levels (the “cold stack”) (Horicuhi et al., 2001).

The correlations of these elements and lake productivity are first reported in Goldberg et al., 2000. The correlation of high ^{10}Be concentrations (even in just the terrigenous fraction) and low ^{10}Be fluxes to the lake with the warm geochemical stack lend further support to the idea that fluctuations in ^{10}Be are primarily controlled by the climatic factors (i.e. glacial erosion and dust deposition) that control sedimentation rate. This set of studies suggests that the increased influx of terrigenous sediment to Lake Baikal the catchment to the lake during glacial periods from decreases the concentration of ^{10}Be in accumulating lake sediments. As with Lake Lisan sediments, changes in atmospheric production rate of ^{10}Be are only evident if they are very large (Horicuhi et al. 2001).

Table 2.1 ^{10}Be concentrations measured by Horiuchi, et al. 1999 from the Academicean Ridge, Lake Baikal. Ages are ^{14}C dates from organic carbon in six stratigraphic horizons (Horiuchi, et al. 1999).

Depth (cm)	Age (yrs)	^{10}Be Bulk Sediment (10^8 A/g)
10-20	--	9.61
20-30	3,500	9.79
30-40	5,900	10.59
40-50	8,300	11.25
50-60	11,000	9.89
60-70	13,600	9.72
72-80	16,100	8.14
80-90	18,100	7.16
90-100	19,800	5.44
100-110	20,900	5.07
110-120	21,900	5.34

Supplement 2D: Summary of Findings by
Belmaker et al. (2008, 2011, 2014)

Lake Lisan reflects the former Pleistocene extent of the Dead Sea. Lisan sediments are widely thought to reflect regional hydroclimate (e.g., Bartov et al., 2002), and sediment cores from this region have been the subject of a variety of paleoclimatic, paleoseismic, and paleomagnetic studies (i.e. Marco et al., 1996; Marco et al., 1998; Machlus et al., 2000). The sediments consist of alternating laminae of aragonite and fine detrital sediments (Stein et al., 1997). Bands of aragonite are thought to precipitate from lake water in response to bicarbonate supplied by freshwater runoff, while bands of detrital sediment, consisting of sand, gravel, and clay represent terrigenous dust from the Sahara and Negev Deserts washed into the lake by water flowing from the Judean Mountains through wadis to the lake (Stein 1997 and Frumkin and Stein, 2001). Some aragonite is also found in detrital layers (Stein et al., 1997 and 2001; Frumkin and Stein, 2004).

Belmaker et al. (2008, 2011, and 2014) utilize these theories to understand the distribution of meteoric ^{10}Be over the past 60 ka of accumulated sediments in the Dead Sea region. Belmaker et al. (2008) measured meteoric ^{10}Be concentrations within a sediment core of Lake Lisan, the former extent of the Dead Sea (Tables 2.2 and 2.3). Ages were determined by U-Th dating of aragonite in the sediments (Tables 2.2 and 2.3) (Hasse-Schramm et al., 2004). After measuring ^{10}Be in the settings of various upland settings that likely contributed sediment to Lake Lisan, the researchers suggest a series of complex hydrologic, climatic, and dust-delivery changes to explain observed variability in meteoric ^{10}Be concentrations. They propose that unless there is a dramatic change in

atmospheric ^{10}Be production rates, variations of ^{10}Be in Lake Lisan sediment reflect hydrological conditions in the watershed and subsequent lake chemistry. Wetter conditions lead to flooding, which results higher flows in wadis, resulting in higher ^{10}Be concentrations in the lake as dust that has settled on upland surfaces is washed into the lake. They also suggest that wet conditions lead to carbonate dissolution. This is important in that they propose that ^{10}Be is preferentially adsorbed to fine silicate particles of airborne dust and flood suspended matter, dissolution of carbonates allows for higher concentrations of these particles, and hence higher concentrations of ^{10}Be (Table 2.2). (Belmaker et al., 2014). During dry periods, little ^{10}Be is delivered to the lake from the environment upland and the accumulation of carbonates in the sediments dilutes the ^{10}Be that is delivered to the lake, resulting in lower ^{10}Be concentrations (Table 2.3). (Belmaker et al., 2014). Overprinting of these hydroclimatic signals only occurs during significant changes in the production rates of cosmogenic nuclides, as occurred during the Laschamp geomagnetic excursion (Belmaker et al., 2014).

Table 2.2 ^{10}Be concentrations in bulk sediment measured by Belmaker, et al. 2008.

Height (cm)	Depth (cm)	Age (yrs)	^{10}Be Bulk Sediment (10^8 A/g)
3505	0	19,470	1.8
2995	510	26,610	1.15
2495	1010	33,610	1.26
2014	1491	40,350	1.21
2010	1495	40,400	1.47
2007	1498	40,440	1.56
1898	1607	41,960	1.27
1895	1610	42,010	1.38
1892	1613	42,050	1.30
1887	1618	42,120	1.44
1883	1622	42,170	1.27
1880	1625	42,220	1.08
1831	1674	48,570	1.30
1825	1680	48,610	1.40
1822	1683	48,630	1.63
1694	1811	49,950	1.64
495	3010	55,200	0.78
5	3500	62,030	1.08

Table 2.3 ^{10}Be concentrations in aragonite sediments measured by Belmaker, et al. 2008.

Height (cm)	Depth (cm)	Age (yrs)	^{10}Be Bulk Sediment (10^8 A/g)
3500	0	19540	0.21
3443	57	20,340	0.23
2280	1220	36,620	0.24
2161	1339	38,290	0.14
2155	1345	38,370	0.07
2150	1350	38,440	0.06
2000	1500	40,540	0.20
1980	1520	40,820	0.59
1970	1530	40,960	0.08
1900	1600	41,940	0.25
1835	1665	42,850	0.16
1654	1846	50,120	0.03
1649	1851	50,150	0.18
1630	1870	50,230	0.12
1500	2000	50,800	0.04
1000	2500	53,000	0.15
5	3495	62,030	0.11

CHAPTER 3

METEORIC ^{10}Be IN BLACKTAIL POND, YELLOWSTONE, AS A RECORDER OF
PALEOENVIRONMENTAL CHANGE

Contribution of Author and Co-Author

Author: Rachel E. Jensen

Contributions: Designed quantitative model framework, applied model to raw data, analyzed data, selected and processed samples in lab, and wrote the manuscript

Co-Author: Jean L. Dixon

Contributions: Designed conceptual model framework, helped design quantitative model framework, oversaw beginning of lab work, discussed results and implications, and edited the manuscript.

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Rachel E. Jensen and Jean L. Dixon
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Abstract

Variations in Meteoric ^{10}Be of accumulated sediments may reflect changing climate and sediment influx. However, relatively few studies have explored how ^{10}Be concentrations reflect paleoenvironmental changes. Previous work showed that measured ^{10}Be concentrations in lake sediments can be successfully modeled by considering sedimentation rates, atmospheric flux, hydrology, and delivery of ^{10}Be to the lake in terrigenous material. Assessment of the relative importance of changes in conditions that influence ^{10}Be concentrations also requires comparison of ^{10}Be concentrations in well-dated cores with independent paleoenvironmental proxies. Here, we provide new insight into the potential for meteoric ^{10}Be in lake sediments to provide paleoenvironmental insights, by exploring a sediment archive of Blacktail Pond in northern Yellowstone National Park. This system provides a record of environmental change that spans the last 15,000 years. We utilize a quantitative model framework predicting meteoric ^{10}Be concentrations with depth in the core and show how modeled concentrations can be used to inform sampling to best capture environmental changes. Measured ^{10}Be concentrations from Blacktail Pond are highly variable and difficult to match with model curves. Though it is possible to model ^{10}Be concentrations with model parameters, we must invoke large changes in ^{10}Be fluxes, which are not supported by existing chemical composition data in the core. Data from this new system highlights the need for quantitative modeling of ^{10}Be concentrations, and suggests more research is needed to understand the disparate controls on concentration of ^{10}Be in lake environments.

Introduction

Lake-sediment records have long been used to reconstruct past climate and environmental characteristics. However, little research has employed sediment cores as archives of geomorphic change. Meteoric ^{10}Be is growing in popularity as a geochemical and geomorphic tool, but it remains underutilized in sediment archives. Due to the unique characteristics of meteoric ^{10}Be , concentrations of this cosmogenic nuclide in accumulated lake sediments have the potential to record environmental change. Specifically, the delivery of meteoric ^{10}Be from atmosphere to the surface in precipitation (Figure 2.1) (Graly et al., 2011) and its strong adherence to fine soils and sediments upon delivery suggest that concentrations of meteoric ^{10}Be in well-dated accumulated sediments provide a record of past climate and sediment influx (Jensen and Dixon, in review).

Interpretations of ^{10}Be concentrations in lake sediments from previous studies vary widely, and often lack a quantitative framework to test the suggested mechanisms. For example, most studies of ^{10}Be in lacustrine sediments suggest changes in ^{10}Be concentrations stem from fluctuations in production rates (Berggren, et al., 2010; Nilsson, et al., 2011; Mann, et al 2012), but the production signal may be obscured by hydrologic and climatic conditions in the lake catchment (Horiuchi, et al., 1999, Horiuchi, et al. 2001, Belmaker, et al., 2008, and Belmaker, et al., 2014). However, when analyzing ^{10}Be concentrations these archives present a suite of challenges, some unique to each system and others common to all ^{10}Be studies. A variety of processes, from terrestrial transport

and soil processes, to lake chemistry and catchment hydrology, will exert influence on ^{10}Be concentrations in accumulated sediments.

Jensen and Dixon (in review) (Chapter 2) propose a quantitative framework for predicting ^{10}Be concentrations in lake sediments and show that even in very different lake environments, measured concentrations can be modeled utilizing relatively simple parameters: sedimentation rate, atmospheric flux, the concentration of ^{10}Be attached to terrigenous sediment, and the fraction of terrigenous sediment in a lake core. The model had been initially tested on datasets from Lake Lisan and Lake Baikal (Jensen and Dixon, in review; Chapter 2). We build on the insights from this study to test the meteoric ^{10}Be system in a new lake archive, one with abundant paleoenvironmental and sediment composition data that constrains model parameters and important environmental changes that influence nuclide concentrations.

Setting

Blacktail Pond is located at just above 2000 m elevation in a kettled glacial outwash terrace. It formed during ice retreat about 15,000 years ago in a remnant meltwater channel (Figure 3.1) (Pierce, 1979). The moraines that underlie the Blacktail Deer Plateau up valley of the Blacktail pond have a cosmogenic age of $14.2 \pm 1.2\text{ka}$, consistent with the Deckard Flats readjustment (Pierce, et al., 2014), and $15.3 \pm 1.4\text{ka}$ down valley (Licciardi and Pierce, 2008), indicating that Blacktail Pond formed between 15,000 and 14,000 years ago (Krause and Whitlock, 2013). The small pond lies in

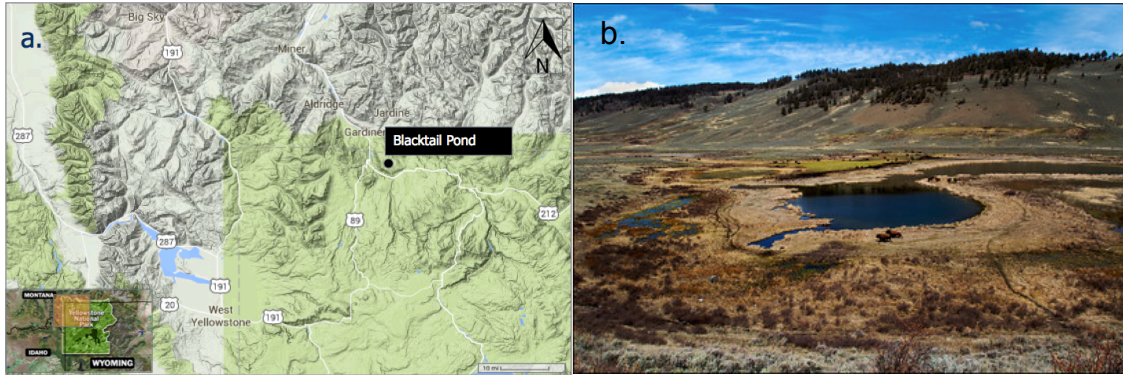


Figure 3.1 a. Blacktail Pond is located in Northern Yellowstone at 44.9555°N, 110.6019°W. b. It is a closed-basin lake located in an abandoned glacial outwash channel.

fine-grained calcareous glacial outwash that supports a sagebrush (*Artemisia tridentata*) steppe. The adjacent gentle slopes are underlain with nutrient rich basalt and andesite, supporting Douglas fir (*Pseudotsuga menziesii*) (Krause and Whitlock, 2013). Other slopes are underlain by nutrient-poor rhyolite support only lodgepole pine (*Pinus contorta*) (Krause and Whitlock, 2013). Seasonal patterns of precipitation in Yellowstone are driven by two large-scale climatic regimes described as “summer-wet” and “summer-dry” (Whitlock and Bartlein, 1993). With high July/January precipitation, Blacktail Pond is located within the summer-wet regime in northern Yellowstone (Whitlock and Bartlein, 1993). Mean annual precipitation is ~39 cm (>100 year average from weather station 8 km from Blacktail Pond, <http://www.wrcc.dri.edu/cgi-bin/cliMAIN.pl?wy9905>).

Blacktail Pond has served as the site of several paleoecological studies that reconstruct the environmental and climatic characteristics of the late-glacial-early Holocene transition in the Great Yellowstone Ecosystem (GYE) from ~14.6 - 6.9ka (Gennett and Baker, 1986; Huerta, et al., 2009; Krause and Whitlock, 2013). This was a

period of rapid environmental change across the globe, featuring rising summer insolation, resulting in rising temperatures, greenhouse gas levels, and the retreat of the large ice sheets and glaciers that dominated the northern regions of North America (Alley and Clark, 1999; Shakun and Carlson, 2010), leading to complex changes in the landscape and the environment (Krause and Whitlock, 2013). The sediments of Blacktail pond, as well as anecdotal records from other lake sediment studies in the region (Krause, et al., 2015; Whitlock, et al., 2012; Whitlock and Bartlein, 1993; Millsaugh, et al., 2004, and others) detail thousands of years of environmental change in the area. These paleoenvironmental studies provide important context for our study of ^{10}Be concentrations in lake sediments, because they constrain several important regional climate and landscapes changes that may similarly be reflected in sediment ^{10}Be concentrations.

Methods

Our work focuses on a sediment core collected in Blacktail pond in 2008, and analyzed by Krause and Whitlock (2013) for pollen, charcoal, and geochemical analysis. Our study has two objectives: firstly to illustrate how quantitative modeling can be used to predict meteoric ^{10}Be concentrations and inform sampling of sediment cores, and secondly to compare model outputs with measured ^{10}Be concentrations to discover the degree to which ^{10}Be concentrations reflect known or inferred changes in paleoenvironmental conditions.

Modeling Blacktail Pond ^{10}Be Concentrations

We utilize a simple numerical model developed in Jensen and Dixon (in review) to evaluate the controls of meteoric ^{10}Be in lake sediments. In this model, the concentration of ^{10}Be was predicted as a function of depth, taking into account both atmospheric and terrestrial ^{10}Be inputs (Figure 2.2). Concentration of ^{10}Be is predicted as a function of depth (Equations 1-3). The flux of ^{10}Be from the atmosphere transports ^{10}Be directly to the lake as wet or dry fallout (Equation 3). Atmospheric flux also delivers ^{10}Be to the environment upland surrounding the lake so ^{10}Be also enters the lake adhered to sediment and dust that is delivered to the lake via erosion or wind (Equation 4). Equation 1 is a combination of both mechanisms. Note that this model was designed with Blacktail Pond in mind. It is a closed basin lake located in a relatively flat basin with gentle adjacent hillslopes. In an open lake, ^{10}Be concentration will be influenced by any inflow and outflow. In a lake with a steep upland environment, runoff must be taken into account (See discussion of hydrologic flux in Chapter 2).

The concentration of ^{10}Be in sediment reflects the total flux of ^{10}Be delivered to the lake and the sedimentation rate, such that:

$$^{10}\text{Be} = \left(\frac{Q_{^{10}\text{BeT}}}{S} \right) \quad (\text{Equation 1})$$

Here, ' $Q_{^{10}\text{BeT}}$ ' ($\text{A}/\text{cm}^2/\text{yr}$) is the total flux of meteoric ^{10}Be to the lake system, and ' S ' ($\text{g}/\text{cm}^2/\text{yr}$) reflects the total sediment accretion rate in the lake (cm/yr) multiplied by sediment density (g/cm^3). Importantly, in this formulation, lake sedimentation acts to dilute ^{10}Be concentrations delivered from the atmosphere. We assume decay is negligible, since the half-life of ^{10}Be is 1.39 My and the sediment archives explored here are

sufficiently young for decay to not significantly influence concentrations. ' Q_{10BeT} ' reflects the atmospheric flux of ^{10}Be directly to the lake and the flux of ^{10}Be in terrigenous material, such that:

$$^{10}\text{Be} = \left(\frac{Q_{10BeA} + Q_{10BeE}}{S} \right) \quad (\text{Equation 2})$$

In this derivation, ' Q_{10BeA} ' (A/cm²/yr) is the flux of ^{10}Be from the atmosphere. If there is no terrigenous input to the lake, the concentration of ^{10}Be reflects only atmospheric input:

$$^{10}\text{Be} = \left(\frac{Q_{10BeA}}{S} \right) \quad (\text{Equation 3})$$

In Equation 2, the flux of ^{10}Be associated with terrigenous sediment (Q_{10BeE}) must be further quantified. ^{10}Be may arrive in the lake via upland erosion (attached to sediment transported in upland streams) or by dust deposition. To account for this input of ^{10}Be to the lake, the amount of allocthonous terrigenous sediment delivered to the lake and the ^{10}Be concentration of that material must be known. Therefore, it is necessary to identify what fraction of the sedimentation rate reflects either terrigenous inputs or in-situ biogenic or chemical sediment production. Accounting for both the atmospherically and terrestrially delivered ^{10}Be to the lake system, the concentration of meteoric ^{10}Be at a given depth in the core is:

$$^{10}\text{Be} = \left(\frac{Q_{10BeA}}{S} \right) + \left(^{10}\text{Be}_E \times \frac{E}{S} \right) \quad (\text{Equation 4})$$

Here, E is the flux of terrigenous sediment (g/cm²/yr), such that E/S reflects the fraction of the total sedimentation rate accounted for by terrigenous sediment input (dust and/or upland erosion products). $^{10}\text{Be}_E$ (A/g) reflects the inherited concentration of ^{10}Be in that sediment.

We developed a suite of models based on our numerical framework to predict ^{10}Be concentrations for the Blacktail Pond core, taking into account varying climatic and environmental scenarios obtained from Krause and Whitlock's 2013 study. We calculated sedimentation rate from the Krause and Whitlock 2013 age-depth model (Figure 3.11) and used the associated geochemical data to estimate terrigenous input. Using the model presented in Graly et al. (2011), we estimated different precipitation-based atmospheric fluxes. Based on the suite of models we developed from raw data, 16 samples were selected from the Blacktail Pond core, spanning different lengths of core, attempting to account for any variations predicted by our range of models. Supplementary Material 3A provides further detailed data and explanation of sedimentation rate calculations and specific parameters used in our model.

Measuring Blacktail Pond ^{10}Be Concentrations

We processed sediment samples for ^{10}Be analysis using a method heavily adapted from Dixon et al. (2016) and von Blanckenburg (1996). Each sample was dried and homogenized. We then subsampled and digested ~1g of bulk sediment in a mixture of hydrochloric, nitric, and hydrofluoric acids (Supplementary Material 3A). Samples were spiked with 600 μg of ^9Be , using a SPEX Be standard. Two procedural blanks were also processed alongside samples. Samples were run through anion columns (analytical grade 100-200 mesh, acetate form) to remove iron and other anionic species. Because of the large amount of calcium carbonate found in the Blacktail Pond sediments, we removed calcium via a sodium hydroxide precipitation before running the samples through cation columns. The precipitate (the calcium fraction) was set aside, while the supernate

(containing beryllium) underwent an ammonium hydroxide precipitation to remove the abundance of added sodium from the sodium hydroxide. The precipitate (containing beryllium) was redissolved and run through cation columns (analytical grade 200-400 mesh, hydrogen form) to remove sodium, aluminum, titanium, and other cationic species, isolating the beryllium. The samples then underwent a second ammonium hydroxide precipitation to form and isolate beryllium hydroxide. This hydroxide was dried down and then oxidized in a muffle furnace to form beryllium oxide. The resulting beryllium oxide powder was packed into targets with niobium to be analyzed via accelerator mass spectrometry at PRIME Lab at Purdue University, Indiana to obtain the ratio of added ^9Be to ^{10}Be in the sediment.

Results: Initial Modeling

Before concentrations of meteoric ^{10}Be were measured in Blacktail Pond sediments, concentrations were predicted by the model of Jensen and Dixon (in review). Predictive models for ^{10}Be concentrations were developed utilizing estimates of past precipitation and terrigenous input based on previously collected pollen, charcoal and geochemical data (Krause and Whitlock, 2013; Pierce et al. 2011). Specific model parameters are provided Supplementary Material 3A. Modeling enabled us to select samples from the core that encompassed possible changes in terrigenous material delivery and climatic conditions and helped us to predict expected concentrations based on realistic parameters.

Sedimentation Rates

An important component of the quantitative model is sedimentation rate, which exerts significant control on ^{10}Be concentrations due to dilution of ^{10}Be fluxes by in-lake sedimentation (Jensen and Dixon, in review). We calculate sedimentation rate (cm/year) for the Blacktail Pond core from the established Blacktail Pond chronology (Krause and Whitlock, 2013). It is important to note that dating of the core section from 759 cm to 810 cm depth, and including our oldest 3-4 samples, was likely underconstrained

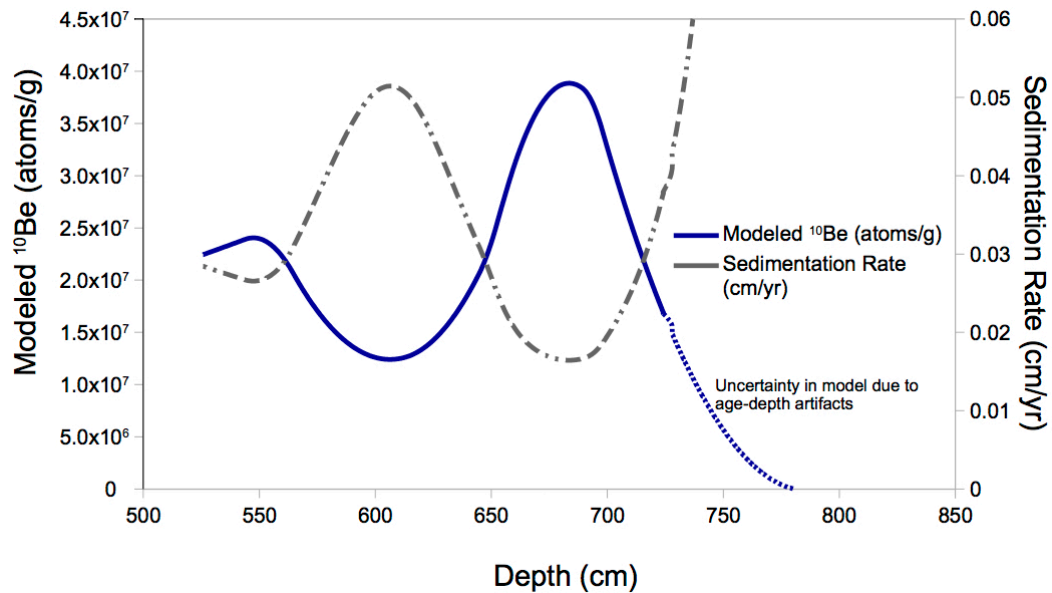


Figure 3.2. Blacktail Pond sedimentation rate (dotted grey) and modeled ^{10}Be concentration (blue), taking into account constant modern precipitation and no terrigenous input. Note uncertainty at beginning of record due to age-depth artifacts.

(Whitlock, personal communication), which influences model results. Sedimentation rates are shown in Figure 3.2. We additionally show one model curve to highlight the inverse patterns between sedimentation rate and ^{10}Be concentrations by this dilution effect (Figure 3.2). This inverse relationship as previously noted by Jensen and Dixon (in

review). Before examining the other aspects of our model framework, it is important to note the calculated sedimentation rates between 750 - 801 cm (corresponding to ages 14,601 -14656 cal yr BP) were anomalous (unusually high or low, even negative) as a result of poorly constrained ages (Figure 3.2).

Precipitation Estimates

We examined the effects of varying precipitation on meteoric ^{10}Be in Blacktail Pond using the model described in Graly et al. (2011) to predict flux based on latitude and precipitation. We used several estimates of precipitation that include the modern estimate of 38 cm/yr, 41.8 cm/yr (10% higher than modern), and 57 cm/yr (50% higher than modern) (Supplementary Material 3A). The effect of these varying precipitation amounts is seen in Figure 3.3 utilizing only Equation 3. Atmospheric flux exerts significantly less influence on the pattern of ^{10}Be concentrations than sedimentation rate (Jensen and Dixon, in review; Chapter 2) and this is evident in Figure 3.3. With changing estimates of precipitation, modeled ^{10}Be concentrations only exhibit small changes in magnitude, while the patterns of concentration are dictated by sedimentation rate (Figure 3.3).

The relationship between predicted ^{10}Be concentrations and sedimentation rate in Blacktail Pond is evident in Figure 3.2. In the model, ^{10}Be concentrations in the Blacktail Pond core are diluted by lake sedimentation unless terrigenous input of ^{10}Be is very high. Our models are based on the idea that meteoric ^{10}Be arrives to a lake environment in one of two ways, in the case of a closed-basin lake like Blacktail Pond. It is delivered directly

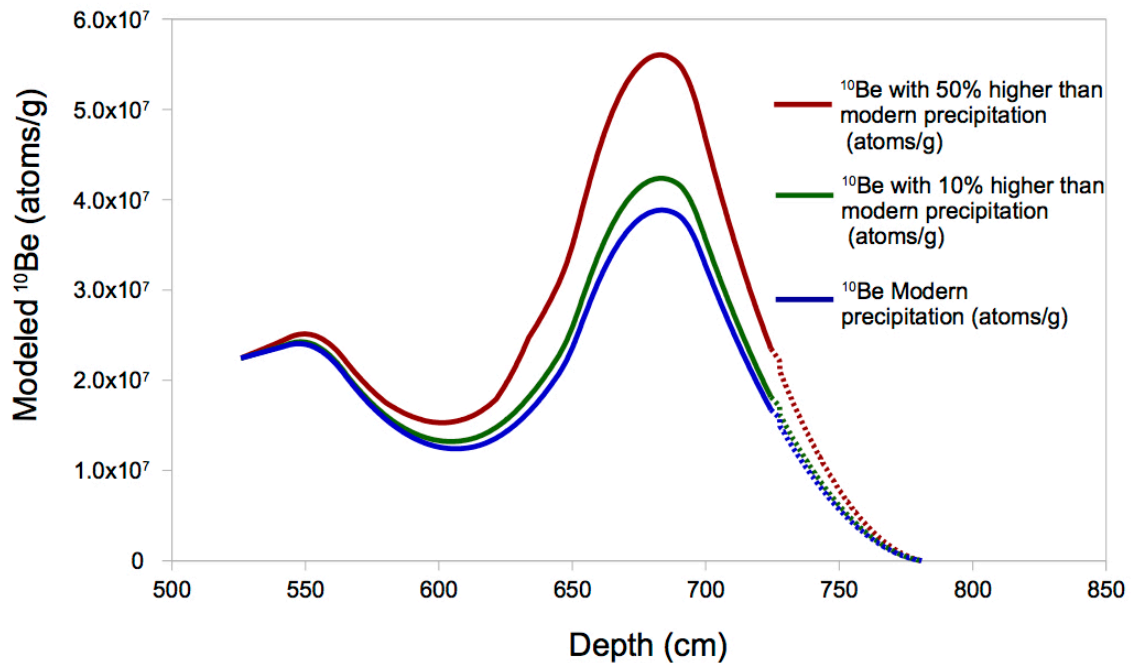


Figure 3.3 Using Equation 3, with varying fluxes obtained from the equation in Graly, et al., 2011, based on differing precipitation estimates, and the sedimentation rate calculated from the age-depth model calculated by Krause and Whitlock, 2013. Concentrations utilizing constant modern precipitation is shown in blue, a precipitation model in which the wettest part of the Holocene is 10% wetter than present is shown in green, and a precipitation model in which the wettest part of the Holocene is 50% wetter than present is shown in red. Uncertain concentrations due to age-depth artifacts are depicted with dotted lines.

to the lake surface from the atmosphere in precipitation or dry deposition. This ^{10}Be is diluted by lake sedimentation. Meteoric ^{10}Be also arrives adhered to fine terrigenous soils and sediments and is delivered to the lake via erosion or as wind-blown dust. This contribution of terrigenous material adds ^{10}Be to lake sediments. Our model results suggest that ^{10}Be concentrations are inversely related to sedimentation rate when there is little or no influx of terrigenous material (Jensen and Dixon, in review; Figure 3.3) The higher the contribution of terrigenous material to the lake, the more ^{10}Be concentration deviates from this inverse sedimentation rate pattern (Figure 3.6).

Estimating In-Situ vs. Terrigenous Sediment Inputs

Using Equation 4, we estimate the concentration of meteoric ^{10}Be at any point along the core based on atmospheric flux (Q_A), the product of sedimentation rate and density (S), the concentration of ^{10}Be in terrestrial sediment $^{10}\text{Be}_E$ (assumed a constant 1×10^8 A/g from Pierce, et al. 2011), and the fraction of terrestrial sediment to total sediment in the core (E/S). We estimated E/S by utilizing titanium XRF data from Krause and Whitlock, 2013 (Supplementary Material 3A). We combined the predicted ^{10}Be concentrations in terrigenous sediments with the concentrations of ^{10}Be predicted by atmospheric flux alone (Equation 4). This suite of models is shown in Figure 3.6, in addition to the models from Figure 3.3, shown for comparison.

Terrigenous input was likely high at the beginning of the Blacktail Pond record, as evidenced from both the records of relative titanium abundance and magnetic susceptibility, both indicators of allocthonous material (Figure 3.4). Abundances of titanium and magnetic susceptibility show similar patterns and indicate high delivery of terrigenous material from 14.6-12 ka, when vegetation cover was sparse and soils were poorly developed (Krause and Whitlock, 2013).

The Ca:Ti data from Krause and Whitlock (2013) also illustrate a transition from deposition of terrigenous material to autochthonous sediment production (Figure 3.5).

Ca:Ti is useful, because carbonate production is used as an indicator of lake

productivity, and hence an estimate of autochthonous material, (Supplementary Material 3B) (Leng and Marshall, 2004), while Ti is an indicator of detrital material (Supplementary Material 3B). In Blacktail Pond, Ca:Ti ratios are higher at the end of

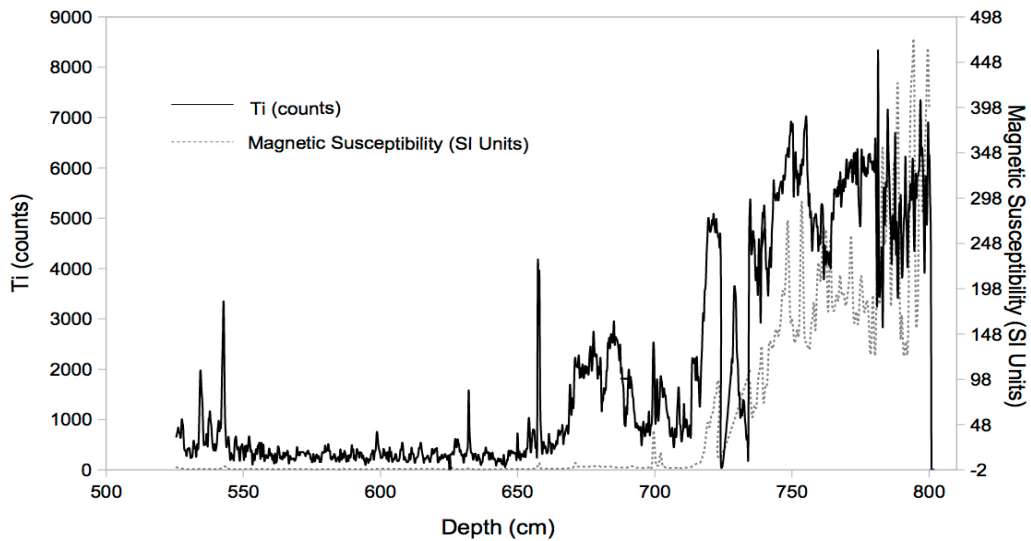


Figure 3.4. Raw Blacktail Pond XRF and magnetic susceptibility data. Relative abundances of titanium are shown in black on the left y-axis and magnetic susceptibility is shown in grey on the right y-axis. Data from Krause and Whitlock (2013).

the record and nearly non-existent at the beginning of the record (Figure 3.5). Low Ca:Ti ratios indicate that less sediment is produced in the lake, while high Ca:Ti indicate more sediment is produced in the lake, based on the conceptual model framework (Figure 2.2), it would be expected that ^{10}Be concentrations would be diluted where Ca:Ti ratios are the highest, and this is not the case. Modeled ^{10}Be concentrations utilize both the relative abundances of titanium (scaled to either 100% or 20% E/S. See Appendix 3A). This set of models produces concentrations that illustrate that the ^{10}Be in Blacktail Pond from 14.6-14.0 ka would likely have arrived in the lake adhered to outwash and dust, depending on what percentage of lake sediment terrigenous material made up (See discussion of E/S in Supplementary Material 3A). After that, as lake productivity

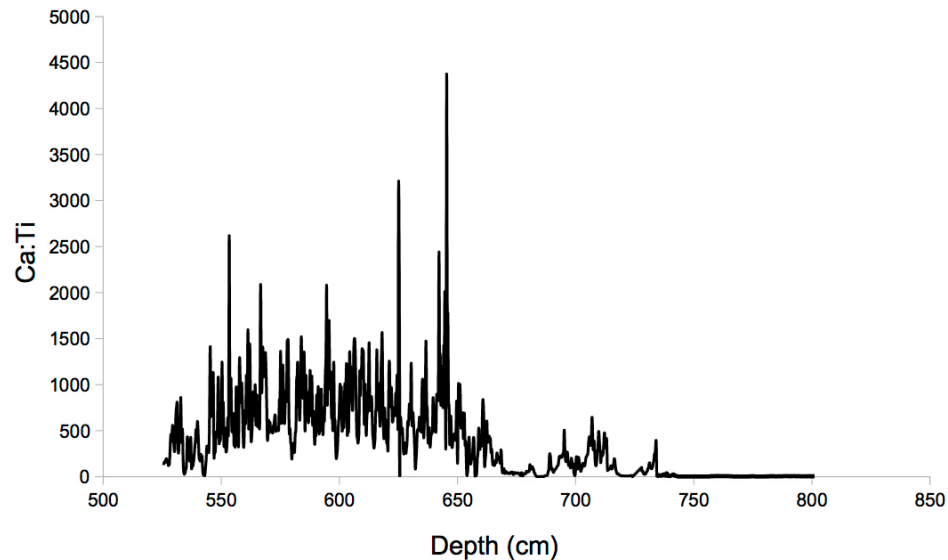


Figure 3.5 Blacktail Pond XRF data. These are Ca:Ti ratios with depth. Data from Krause and Whitlock (2013).

and landscape stability increased, the ^{10}Be in the lake would primarily have arrived directly to the lake via atmospheric flux, where it would have been diluted by lake sedimentation. Remember, the effect of precipitation in our models is minimal using flux derived from Graly, et al. (2011) (Jensen and Dixon, in review; Figure 3.3). Changing the magnitude of flux does not change the pattern of ^{10}Be concentration with depth; it changes the magnitude of ^{10}Be concentration, especially when sedimentation rate is low (Figures 3.2 and 3.3).

Based on the models utilizing terrigenous input (Equation 4), ^{10}Be concentrations are largely predicted by sedimentation rates unless terrigenous input is high, as it is at the end of the record (seen in Figure 3.3). As the percentage of terrestrial sediment increases, the predicted ^{10}Be concentrations deviate from those predicted purely by dilution by

sedimentation rate. Note how the models differ at the lowest depths in the core, when terrigenous input was at its highest (Figure 3.6).

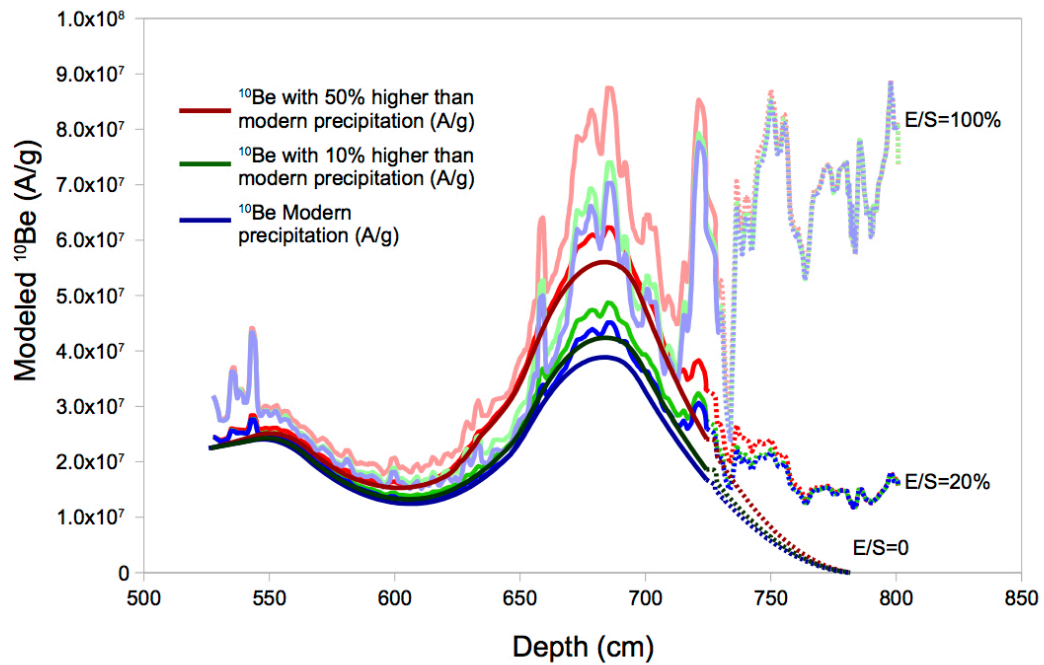


Figure 3.6 Models with different estimates of precipitation (modern precipitation in blue, 10% wetter than present in green, and 50% wetter than present in red) and different estimates of the fraction of terrigenous sediment in the core. The lower curves utilize a E/S (Equation 4 in Appendix 3A) that is 20% at the most and the upper curves utilize a E/S that is 100% at the most. The lowest curves are the flux-only curves from Figure 3.3.

Results: Analysis of meteoric ^{10}Be in Lake sediments

Measured ^{10}Be concentrations

Measured ^{10}Be concentrations range from 2.62×10^7 A/g to 3.39×10^8 atoms/g, averaging 9.33×10^7 atoms/g ($\pm 3.47 \times 10^6$ atoms/g (Table 3.1), with significant variability across the core (Figure 3.7). ^{10}Be concentrations were blank corrected by subtraction (average $^{10}\text{Be}/^9\text{Be}$ of two chemical processing blanks = $6.4 \pm 0.63 \times 10^{-15}$).

PRIME Lab reported that samples M6, M7, and M15 showed very low current, and so are subject to more error.

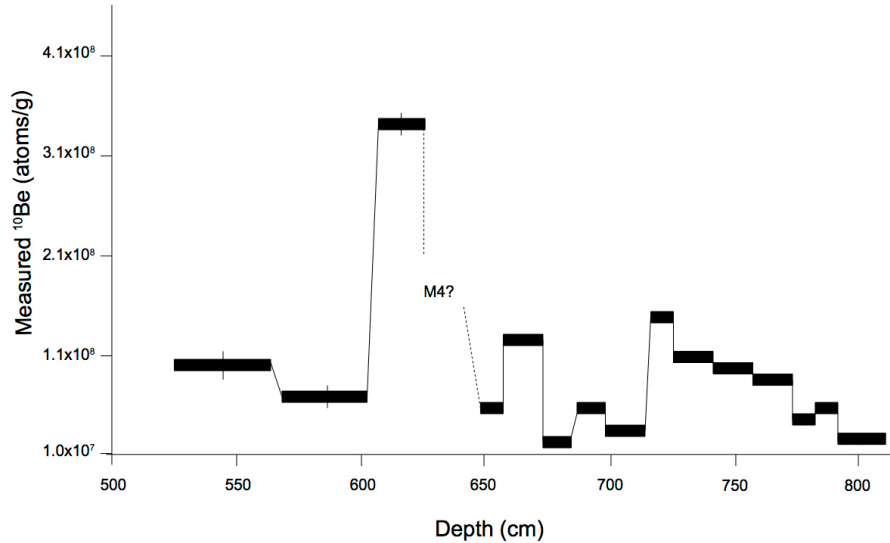


Figure 3.7 Measured concentrations of ^{10}Be in bulk sediments of the Blacktail Pond core. Black bars span the sample depth intervals and vertical error bars reflect analytical error and blank uncertainty. Most vertical error bars are small enough to be obscured by the symbols. Sample M4 was lost in processing.

With the exception of the bottom of the core, we find little systematic variation in meteoric ^{10}Be concentrations across the sampling interval. Starting at the bottom of the core (~ 800 cm depth), ^{10}Be concentrations increased gradually from M16 to M10, reflecting increasing concentrations during the early Holocene until ~ 14 ka (Figure 3.7). However, we observe a decrease in concentrations at sample M9 (~ 13.4 ka), where ^{10}Be concentrations were nearly as low as at the beginning of the record. Samples near the top of the core (M1-M7; 7.6 -11.9 ka) in general showed higher variability compared to samples at the bottom of the core (M8-M16; 12.7-14.7 ka).

Table 3.1. Samples from the Blacktail Pond Core and their corresponding, depths, ages, $^{10}\text{Be}/^9\text{Be}$ ratios and measured concentrations.

Sample Name	Depth (cm) [Average]	Age (yrs) [Average]	$^{10}\text{Be}/^9\text{Be}$ (10^{-15})	^{10}Be Concentration (atoms/g)
M1	525-563 [544]	6985-8248 [7667]	1261.94	1.00×10^8
M2	567-604 [585]	8476-9324 [8900]	782.16	6.65×10^7
M3	607-625 [616]	9401-9769 [9585]	4161.70	3.39×10^8
M4*	625-643 [634]	9769-10245 [10007]	--	--
M5	647-657 [652]	10245-10773 [10509]	765.56	5.84×10^7
M6**	657-673 [665]	10773-11621 [11197]	1590.02	1.30×10^8
M7**	673-683 [678]	11621-12222 [11922]	335.00	2.62×10^7
M8	687-697 [692]	12404-13048 [12726]	693.33	5.81×10^7
M9	697-713 [705]	13048-13771 [13410]	457.38	3.82×10^7
M10	717-725 [721]	13909-14121 [14015]	1838.12	1.49×10^8
M11	725-741 [733]	14121-14450 [14286]	1413.43	1.12×10^8
M12	741-757 [749]	14450-14601 [14625]	1239.89	9.87×10^7
M13	757-773 [765]	14601-14649 [14631]	1081.85	8.86×10^7
M14	773-781 [777]	14649-14652 [14651]	620.54	4.76×10^7
M15**	781-789 [785]	14652 [14652]***	759.42	6.07×10^7
M16	789-810 [800]	14652-14656 [14654]	363.07	2.65×10^7

*M4 lost during processing

**These samples had very low current

***The sedimentas associated with this sample are poorly dated

Comparison of Measured Concentrations to Modeled Concentrations

Measured concentrations vary notably from modeled values, though average concentrations occur roughly at the same magnitude (Figure 3.8). Several samples fall far from model predictions. M3, the sample with the highest measured ^{10}Be concentration, is an order of magnitude higher than any modeled predictions for that depth interval (Figure 3.8). M7, the sample with the lowest measured ^{10}Be concentrations, is lower than all modeled predictions for that depth interval (Figure 3.8). M7 concentrations are ~30% lower than models that are based solely on atmospheric flux (with no terrestrial input), and ~70% lower than the modeled concentrations that maximize terrigenous influx (Ti-

based E/S, where the highest E/S value is scaled to 100% and carbonate-based E/S).

These differences seem most apparent at higher depths in the core (younger ages), where samples do not fall within the range of models.

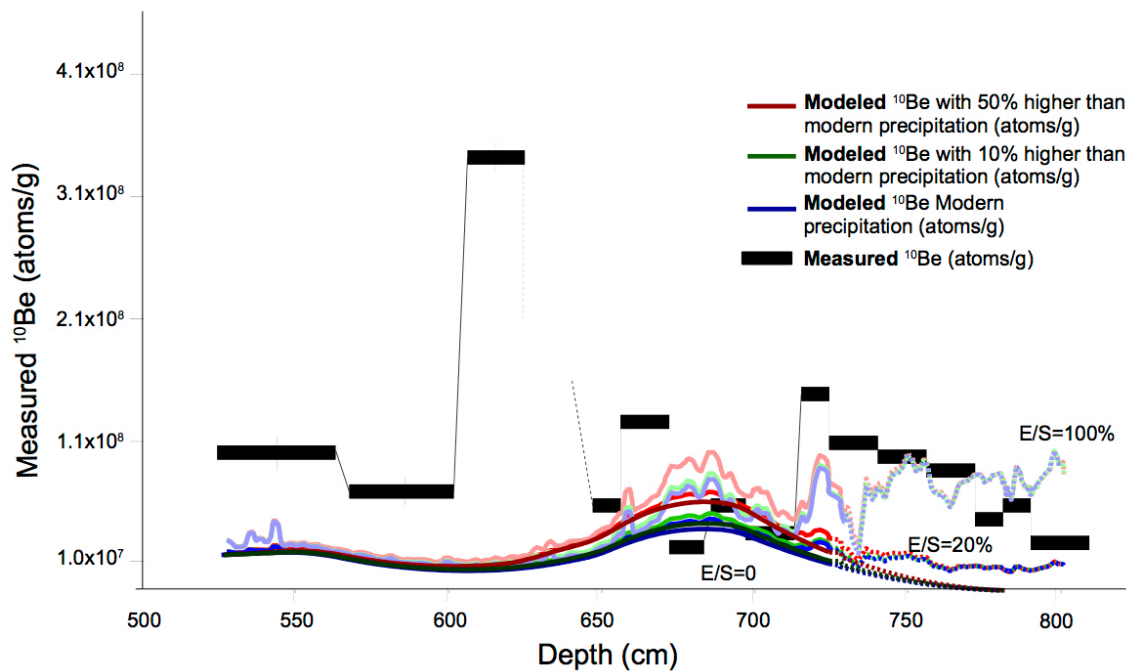


Figure 3.8 Measured concentrations (black bars) plotted with a range of selected models developed for Blacktail Pond. These models are shown in Figure 3.6.

Since measured concentrations do not correlate closely with modeled concentrations, measured concentrations were additionally compared directly to existing proxies for environmental change, such as oxygen isotope records, carbon, pollen, charcoal, and geochemical data (Krause and Whitlock, 2013), as well as our own measurements of carbonate percentage. These comparisons are detailed in Appendix 3C. Of special interest are the comparisons of ^{10}Be to Ti, Ca, Sr, Se, Hg, As, S, and stable oxygen isotopes.

Discussion

The measured concentrations of ^{10}Be from Blacktail Pond can be interpreted through our conceptual model framework (Figure 2.2). Because of the dominant control of sedimentation rate in the model, measured concentrations were initially compared to sedimentation rates (Figure 3.9). Unlike our modeled predictions (e.g., Figures 3.2 and 3.3), measured ^{10}Be concentrations do not show a clear inverse relationship to sedimentation rate. Instead, measured concentrations appear to correlate with sedimentation rate, especially after 14 ka (M10). From its formation at 14.6ka, Blacktail Pond experienced a high influx of glacial outwash material and/or dust, and so sedimentation rates were likely quite high and decreased with time. However, dating the record during this time was difficult. The oldest AMS ^{14}C age obtained by Krause and Whitlock (2013) was 14,549 cal yr BP at 786.26 cm. The age-depth model used in that study produced an age of formation 14,650 cal yr BP, which, as they point out, is consistent with the age of recessional moraines dated by Licciardi and Pierce (2008). If sedimentation rates were high at the beginning of the record and decreased to 14 ka, then it is only at the beginning of the Blacktail Pond record that measured ^{10}Be concentrations show an inverse correlation to sedimentation rate, perhaps illustrating a change in processes governing ^{10}Be concentration. By comparing ^{10}Be concentrations in this portion of the core to modeled ^{10}Be concentrations, there are similarities (Figure 3.8), especially when terrigenous influx is considered. Importantly however, no modeled concentrations are clear matches for the magnitude of measured concentrations, there are similarities in the trends of ^{10}Be concentration. This trend is most evident when Ti-based

E/S is scaled to 20%, a moderate estimate of terrigenous influx (Figure 3.8). This trend suggests that while there was high influx of ^{10}Be -bearing terrigenous material to the lake during this time, it was not enough to completely obscure the control of sedimentation rate on ^{10}Be concentrations.

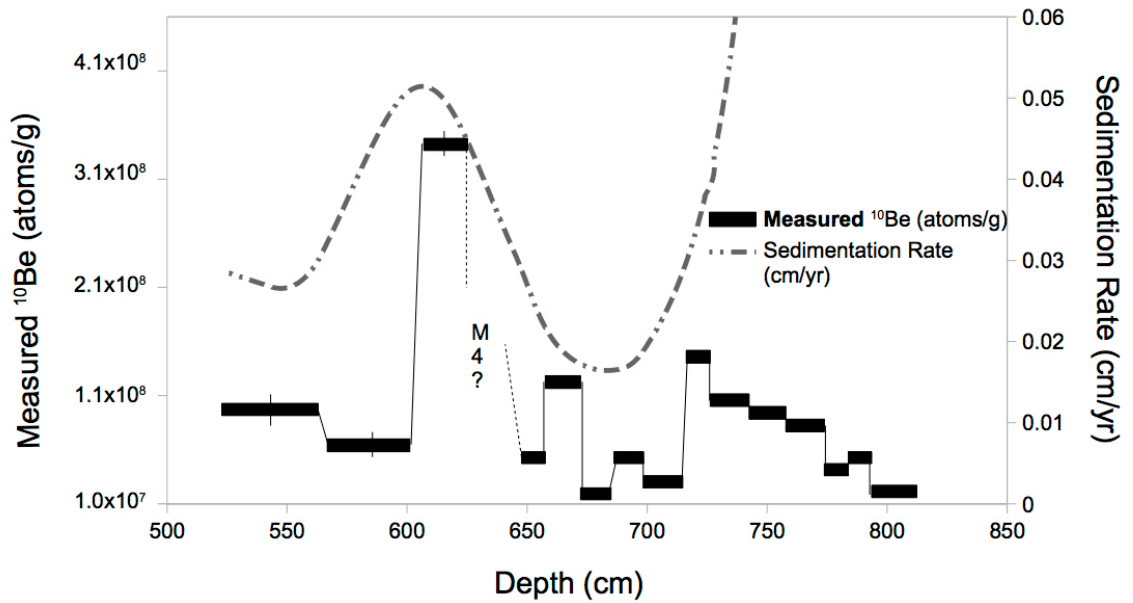


Figure 3.9. Measured ^{10}Be concentrations (black bars) on left y-axis and sedimentation rate (grey dotted curve) on right y-axis

After 14 ka, lake sedimentation does not appear to have a dilution effect on ^{10}Be concentrations. Rather, higher sedimentation rates instead correspond with higher ^{10}Be concentrations (Figure 3.9). Since the sedimentation in Blacktail Pond after 14 ka is dominated by non-detrital, authochthonous material; the biggest contribution of ^{10}Be to the lake was from atmospheric flux, rather than terrigenous material. For example, we can consider the unexpectedly high concentration of M3. If no additional terrigenous input is assumed, direct atmospheric flux would have to be about $10^7 \text{ A/cm}^2/\text{yr}$ for

modeled ^{10}Be concentrations to match the concentration of M3 (~9.5 ka). This atmospheric flux is more than an order of magnitude higher than any flux predicted for the region by the Graly, et al. (2011) model (averaging 8.5×10^5 atoms/cm²/yr). The average global flux of ^{10}Be to the surface is around 1.21×10^6 atoms/cm²/yr (Monhagen et al. 1986; Masarik and Beer, 2009), close to the fluxes produced by the Graly, et al. (2011) model. To account for atmospheric flux as calculated by Graly, et al. (2011), precipitation would have to be more than 10 times higher than the modern annual precipitation of 38 cm/yr in Yellowstone. While sample M3 encompasses what is likely the wettest and warmest part of the record during the solar insolation maximum at (Krause and Whitlock, 2013), an annual precipitation regime of 380 cm/yr is higher than most precipitation rates in modern tropical rainforests (Ritter, 2006)! According to the paleoenvironmental records from Blacktail Pond, the surrounding area supported Douglas fir and lodgepole pine at 9.5 ka (Krause and Whitlock, 2013). Records from the nearby lakes support the idea of cool winters, warm summers, and wetter than modern conditions (Krause and Whitlock, 2013; Whitlock 2012), not a rainforest.

Terrigenous input must also be considered to understand the measured concentrations of samples like M3. Ti counts served as an estimate of the percentage of terrigenous sediment in the core (Supplementary Material 3A), but a constant estimate of the ^{10}Be concentration on terrigenous sediment (1×10^8 A/g from Pierce et al. 2011) was assumed for the entire record. It is unlikely that the concentration of ^{10}Be remained the same over this 7 ky record due to changes in the deposition of ^{10}Be to surface and sediment transport. If the measured concentration of M3 is realistic and all of the other

model parameters remain the same, the concentration of ^{10}Be on terrigenous sediment would have to be 7.39×10^9 A/g, which is quite high, especially if samples M2 and M5 have much lower ^{10}Be concentrations.

Hydrologic ^{10}Be sources to the lake system should also be considered, though Dixon and Jensen (in review) suggest these are minimal in most lake environments due to retainment in soils. While Blacktail Pond is a closed basin lake today, such that the only inputs of ^{10}Be are from the atmosphere and dust sources, it is possible that there are hydrologic effects on the ^{10}Be concentrations in the Blacktail Pond record. In Chapter 2, we include the possibility of a hydrologic contribution of ^{10}Be to the sediment record by integrating the atmospheric flux across the catchment area. If atmospheric flux is uniform across the catchment, then atmospheric flux is simply multiplied by the ratio of the catchment area to the lake surface area (Equation 4). To calculate ^{10}Be concentration, this flux is simply added to the atmospheric flux and the flux of ^{10}Be in terrigenous material before being divided by sedimentation rate (Chapter 2; Jensen and Dixon, in review). If this addition is invoked to explain the high ^{10}Be concentrations of M3, assuming a uniform modern flux, the catchment area would have to be approximately 21 times larger than the lake. Previous quantitative examination of ^{10}Be concentrations in lake systems (Jensen and Dixon, in review), explored regions that were much larger (e.g., Lake Baikal and Lake Lisan had catchment areas $\sim 18\times$ and $16\times$ larger than the respective lakes). Blacktail Pond is comparatively very small, currently covering about $7,400 \text{ m}^2$, compared to Lake Baikal and Lake Lisan, at $31,700 \text{ km}^2$ and $2,500 \text{ km}^2$, respectively. A contributing area of $155,400 \text{ m}^2$ seems reasonable. However, assuming constant and

uniform modern flux, to match ^{10}Be concentrations of M5, about 900 years prior, the catchment area must be close to the same size as the lake area during the M5 interval. This means that in the 900 years between M5 and M3, if the catchment area remained about the same, the lake would have to become 21 times smaller in M3 than in M5. Similarly, Blacktail Pond would have had to grow by at least a factor of 10 from M3 for ^{10}Be concentrations to match those measured in M2 (Table 2.1). For this to be true, M3 would have to represent an especially dry interval, compared to M2 and M5. The $\delta^{18}\text{O}$ record for Blacktail Pond represents summer evapotranspiration, driven by summer temperatures and water balance. This means that $\delta^{18}\text{O}$ concentrations are higher during periods of high evaporation, when lake size decreases. Because the ^{10}Be and $\delta^{18}\text{O}$ records share peaks (Figure 3.10), it may be possible that the larger difference between catchment area and lake area can account for high ^{10}Be concentrations. However, there is significant uncertainty that these required shifts in lake size are possible for this system. The sensitivity of our model to changes in catchment area should be reexamined.

Sample M7 (See Table 3.1) also stands out in that its ^{10}Be concentration was low. Based on the relative abundances of titanium in the core (Figure 3.4), terrigenous input was not unusually low; in fact, there is a peak in titanium at that time. Sedimentation rates were low, which would theoretically increase ^{10}Be concentration. The concentration of M6 was higher than that of M7, though sedimentation rate increased after the M7 interval. If the Graly, et al. (2011) flux are adjusted to account for the low ^{10}Be concentration, precipitation must have been 13cm/yr lower than modern precipitation during that interval. While the early Holocene (M7 spans 11621-12222 ka) is

characterized as being cool and dry, with increasing temperature and precipitation (Krause and Whitlock, 2013) it is unclear whether or not 13cm lower than modern precipitation is likely. For M7 the concentration of ^{10}Be on terrigenous sediment cannot even be considered all other model parameters remain the same. Even if incoming terrigenous sediment does not have any ^{10}Be adhered to it, modeled concentrations cannot decrease enough to match the measured concentration of M7, because the measured concentration of M7 is below any modeled concentration for sedimentation rates during that period. Additionally, contributing area alone cannot be invoked to explain concentrations, since based on our model, even a contributing area to lake area ratio of 0 would not produce a sufficiently low ^{10}Be concentration to match the measured ^{10}Be concentration of M7. Here, another mechanism must be involved, or we must attribute this low concentration to loss of sample in the laboratory. Laboratory error is discussed further below.

When considering the rest of the ^{10}Be record, it is evident that sedimentation rate does not exert the control that was expected, with the possible exception of the first 600 years of the record (14.6-14.0 ka). Samples M8, M9, M14, M15, and M16 are within the bounds of modeled predictions. Otherwise, there are a number of points of interest to consider. First, the ^{10}Be concentrations of M1, M2, M5, M6, M10, M11, M12, and M13 can be matched by changing the concentration of ^{10}Be on incoming terrigenous sediment by less than an order of magnitude. This means that many of the measured concentrations can be accounted for by our model. However, the concentration of ^{10}Be on terrigenous sediments is the product of complex climate and landscape mechanisms,

ranging from changes in atmospheric flux to sediment transport. The deviations of measured concentrations from modeled concentrations can point towards changes in environmental conditions that influence ^{10}Be in a lake system. Given the other paleoenvironmental data from Blacktail Pond (climatic factors or terrigenous input), it is still hard to justify why the concentration of ^{10}Be on incoming detrital sediments would change in the ways it would have to in order to account for all concentrations.

Second, there is surprisingly similarity to both our carbonate measurements (Supplementary Material 3C) and the total carbon, carbonate, and Ca:Ti measurements from Krause and Whitlock (2013) (Supplementary Material 3C), in that carbon and carbonate is high when ^{10}Be is high. This correlation was unexpected, given that these are indicators of lake-produced sediment, which would dilute ^{10}Be concentrations. This relationship is reinforced by the relationship of ^{10}Be concentration to sedimentation rate. Additionally, there is some correlation between ^{10}Be concentration and relative abundances of Sr, Se, Hg, As, and S (Supplementary Material 3C). This and the relationship of ^{10}Be to carbon may be a product of one or several chemical processes in the lake that we have not considered. Most intriguingly, there may be a relationship between ^{10}Be and $\delta^{18}\text{O}$ (Figure 3.10). While the changing ratio of contributing area to lake area might be a factor, the relationship requires further examination.

Finally, the possibility of laboratory error cannot be ruled out. There is no standard procedure for processing of lake sediments for ^{10}Be . This means that our procedure was adapted from methods designed for extracting ^{10}Be from terrigenous soils (Dixon et al. 2016 and von Blanckenburg 1996). We encountered a variety of

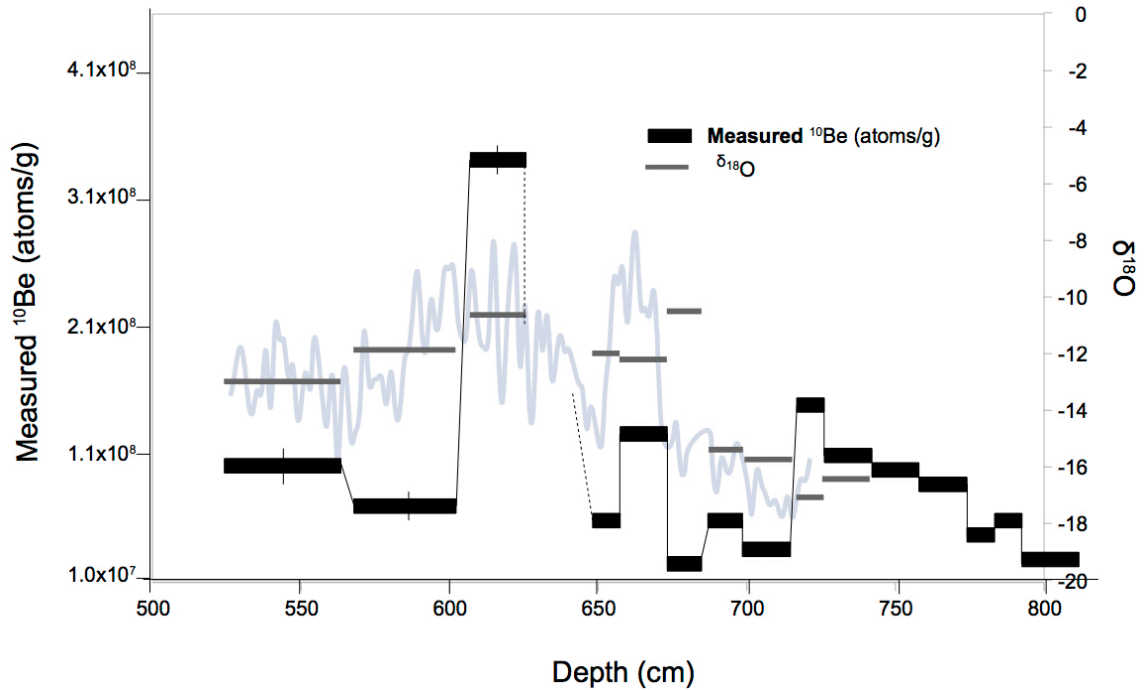


Figure 3.10. Measured ^{10}Be concentrations (black bars) on left y-axis. $\delta^{18}\text{O}$ on right y-axis. Binned $\delta^{18}\text{O}$ values by sample are depicted as grey bars, while raw values are depicted in the background.

complications. The abundance of CaCO_3 in Blacktail Pond sediments meant that samples had to undergo NaOH precipitation to remove the high levels of calcium before $^{10}\text{BeOH}$ could be precipitated in ammonium hydroxide, as per the terrigenous ^{10}Be protocols (Supplementary Material 2A). Every additional step allows for error and sample loss. There is also possibility that despite repeated digestions in a variety of acids, some ^{10}Be might have remained bound within undigested material before the addition of ^9Be . There are no certified standards of ^{10}Be in lake sediments, which would help to validate that the laboratory methods developed specifically for this study were robust. Additionally, while prior blank measurements indicated that the lab was suitable for meteoric ^{10}Be processing, anecdotal evidence suggests it was not a homogeneously clean or dust-free environment. We observed significant spatial and temporal heterogeneity in settling dust

in the lab environment, especially considering significant dry down time occurred in an exhaust hood that lacked air filtration. Therefore, it is possible that despite adequate measurements in our two processing blanks, the lab environment resulted in sample contamination. While samples are dried down together on a hot plate in the hood, sample vessels are open for hours at a time, allowing not only for contamination from the environment, but between samples, especially if samples are heated vigorously. Future work should ensure adequate air filtration for consistency of the lab environment and reduce open vessel time.

Conclusion

The measured concentrations of meteoric ^{10}Be in the Blacktail Pond record were both higher than expected and were not diluted by sedimentation rate. Our model framework predicted that ^{10}Be concentrations would reflect atmospheric addition of meteoric ^{10}Be and its dilution by lake sedimentation, unless the lake experienced such a high influx of ^{10}Be -bearing terrigenous material that the sedimentation rate signal was obscured. This may be the case at the beginning of the Blacktail Pond record, immediately after glacial recession, when there was little vegetation cover and much of the sediment in the lake was derived from terrestrially delivered dust and glacial outwash. Uncertainties in sedimentation rate during this period make it difficult to test this assumption. As for the rest of the record, it is possible to explain all but the lowest ^{10}Be concentrations by changing the concentration of ^{10}Be on incoming terrigenous sediment or modestly increasing the catchment area. The lowest concentrations cannot be accounted for by changes in ^{10}Be on terrigenous sediment, and so would require a

dramatic increase in lake size. A change in lake size might also explain the possible correlation of high ^{10}Be concentrations with high values of $\delta^{18}\text{O}$. To test this relationship will require ^{10}Be from different lakes with $\delta^{18}\text{O}$ records. Our previous work indicated that our model framework could account for meteoric ^{10}Be concentrations in vastly different environments, but the lack of clear agreement between modeled and measured concentrations at Blacktail Pond suggest that more testing of our model is needed. ^{10}Be concentrations in Blacktail Pond may be especially sensitive to small changes in model parameters (especially the concentration of ^{10}Be on terrigenous sediment) because of its small size. Small lake and catchment size may result in highly variable concentrations because small environmental changes to either the lake or upstream environment could potentially result in large changes in the nuclide signal of lake sediments since signals will have little ability to be buffered out. Future work should continue to explore meteoric ^{10}Be in a variety of lake records using our quantitative model to better understand the behavior of and controls on ^{10}Be in lacustrine environments. Meteoric ^{10}Be concentrations in sediment archives remain an intriguing possible paleoenvironmental proxy, though more testing is needed.

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Supplementary Material

Meteoric ^{10}Be in Blacktail Pond as a Recorder of Paleoenvironmental Change in Yellowstone

Supplement 3A: Methods

Supplement 3B: Estimating In-Situ vs. Terrigenous Sediment Inputs

Supplement 3C: Comparisons of Measured Concentrations to known paleoenvironmental proxies

Supplement 3A: Detailed Methods

Modeling Predicted Concentrations. We developed a simple, quantitative model to evaluate controls on meteoric ^{10}Be in lake sediments by presenting a numerical model that predicts the concentration of ^{10}Be as a function of atmospheric flux, sedimentation rate, and terrestrially derived ^{10}Be (See Chapter 2).

We utilize the age-depth model from Krause and Whitlock (2013) (Figure 3.11). From this chronology, we calculated sedimentation rate (cm/year) for the Blacktail Pond core by dividing change in depth by change in time. Eight ages were obtained from AMS radiocarbon dating of plant macrofossils, charcoal, and pollen concentrates, as well as age layers from age-constrained volcanic eruptions (Krause and Whitlock, 2013). Calendar ages were calculated from ^{14}C ages with CALIB 6.0 software and the age-depth model was obtained from MCAgeDepth software (Krause and Whitlock, 2013). There are calculated 1092 ages for 0.25cm intervals

In order to calculate atmospheric flux ($Q_{^{10}\text{BeA}}$) (Equations 2-4), we utilize the model presented by Graly et al. (2011). However, any estimated flux for the region could be inserted as the flux term $Q_{^{10}\text{BeA}}$. Flux is calculated as a function of latitude and annual precipitation, thus allowing for estimation of ^{10}Be concentration under changing climatic conditions. The latitude $44.955369^\circ \text{ N}$ is, of course, constant, but it is likely that precipitation changed over the course of the Blacktail Pond record. We assign changing precipitation values based on qualitative estimates from paleoclimate studies of the region and the site itself (Huerta, et al., 2009; Krause and Whitlock, 2013; Pierce, et al.,

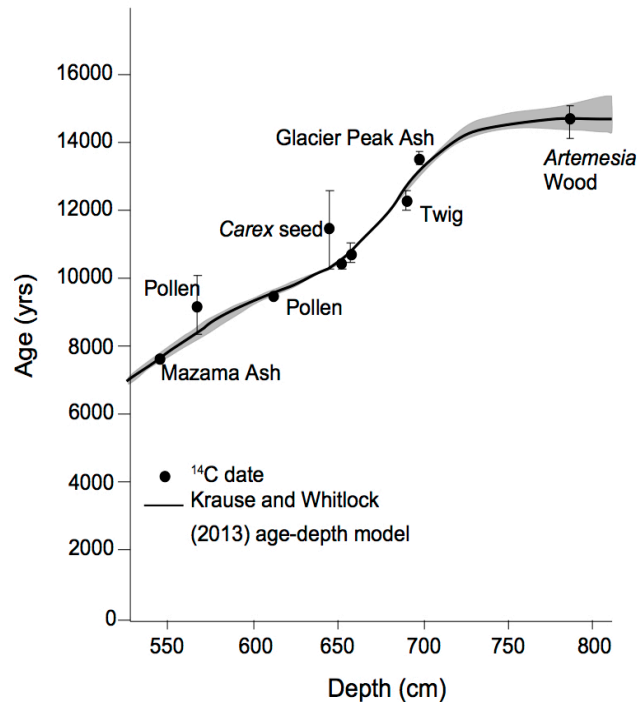


Figure 3.11. Age-depth model for Blacktail Pond based on radiocarbon and tephrochronology. Adapted from Krause and Whitlock (2013).

2014) (Figure 3.12). We develop a model assuming constant modern precipitation, a model in which the wettest part of the record, around 9000 cal yr BP, was 10% wetter than today, and in which it was 50% wetter (Figure 3.12). The precipitation regime for the area during the late glacial-early Holocene transition is characterized qualitatively by Krause and Whitlock, 2013, Huerta, et al., 2008; Bartlein, et al., 1998, Mock and Bartlein, 1995, Whitlock and Bartlein, 1993, and others. Since Blacktail Pond is located in the summer-wet area of Yellowstone, the summer insolation maximum occurred around 9ka (Krause and Whitlock, 2013), and winter precipitation is expected to have been high 9ka (Whitlock, et al., 2012), the highest annual precipitation in the record probably occurred then. It is important to note that the summer-wet/summer-dry

conceptualization may not describe precipitation patterns at Blacktail Pond thoroughly, but changes in precipitation, unless they are quite dramatic do little to affect patterns of ^{10}Be concentration (Figure 3.3; Chapter 2). We found that utilizing constant modern values for precipitation versus estimating modest changes in precipitation does little to change patterns of ^{10}Be concentrations in the modeled outputs.

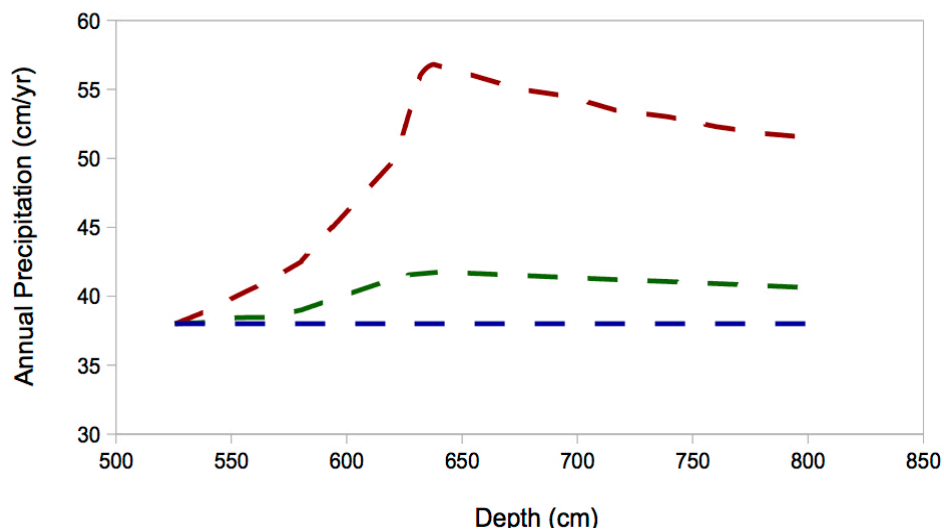


Figure 3.12 Precipitation models assuming that conditions at the end of the record include constant modern precipitation at Blacktail Pond (38cm/yr) and the wettest part of the Holocene, at 9 ka, was 50% wetter (red curve) or 10% wetter (green curve) than modern. Constant modern precipitation is shown in blue. These precipitation inputs were used to calculate flux using the Graly, et al. (2011) model resulting in the ^{10}Be concentrations shown in Figures 3.3 and 3.6. We also used models that assumed constant precipitation values (38 cm/yr, 41.8 cm/yr, and 57 cm/yr). These resulted in fluxes, and therefore ^{10}Be concentrations, nearly indistinguishable from those shown in Figures 3.3 and 3.6.

Estimation of ^{10}Be from terrigenous sediment involves two components: the concentration of ^{10}Be attached to terrigenous sediment ($^{10}\text{Be}_E$) and the fraction of terrigenous sediment in the core (E/S) (Equation 4). For $^{10}\text{Be}_E$, we assumed a constant

value of 1×10^8 A/g, estimated from average ^{10}Be concentrations on loess from the region (Pierce et al., 2011). E/S was first estimated from XRF data collected by Krause and Whitlock, 2013 (Supplementary Material 3B). We scale the counts of Ti to estimate the percentage of terrigenous material at any point in the core. We developed two models, one in which the highest abundance of Ti, which occurred approximately 14.6 years ago, represents 100% of the sediment in the core as from terrestrial sources, rather than produced in the lake (i.e. $E/S=1$), and the other in which the highest abundance of Ti represented 20% of the sediment in the core as from terrestrial sources (i.e. $E/S=0.2$) (Figure 3.6). We also calculated E/S based on carbonate content, rather than XRF data (Figure 3.13). We used the relative abundance of titanium to estimate terrigenous input. Since Ca:Ti ratios are commonly used to characterize the conditions that lead to authigenic carbonate precipitation, carbonate content may be used to estimate the amount of sediment produced in the lake. By this method, $E/S=1$ -Carbonate Fraction.

Carbonate fraction for each sample was measured by digesting sample in low molarity hydrochloric acid and titrating the reacted solution with sodium hydroxide (NSF Environmental Chemistry Library). This allows for a rough calculation of the amount of carbonate that reacted with the acid and, hence, the percentage of the sediment made up of carbonate. Additionally, the unreacted sediment was weighed to determine the amount of material lost. These new values were input for E/S to create a second suite of models (Figure 3.13).

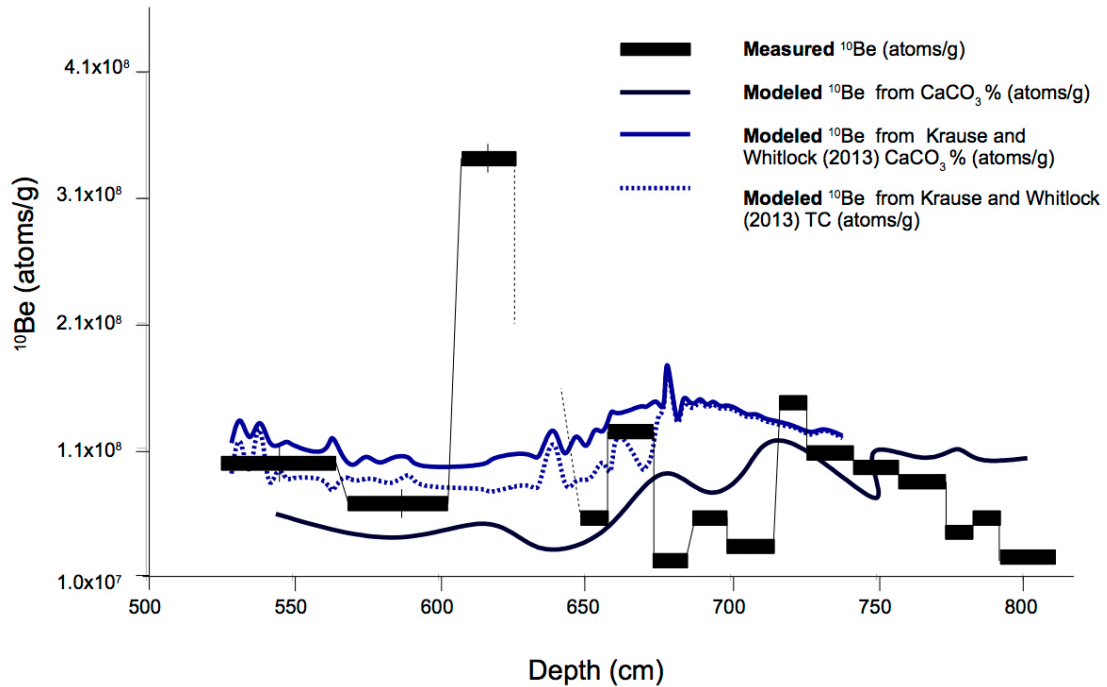


Figure 3.13. Measured concentrations of ^{10}Be (black bars) and modeled concentrations (blue), using terrigenous input (E/S) derived from carbonate percentages. Bottom-most model is based on our laboratory CaCO_3 measurements, top-most model is based on CaCO_3 measurements from Krause and Whitlock (2013), and middle model is based on total carbon measurements from Krause and Whitlock (2013).

Sample Selection. The range of models taking into account varying precipitation and erosion rates predict the concentrations of ^{10}Be as a function of depth in the Blacktail Pond core and allowed for sample selection. Based on the suite of models we developed we selected 16 samples, spanning different lengths of core, attempting to account for any variations predicted by our range of models (Table 3.1).

We took the selected samples from the sample core described by Krause and Whitlock, 2013. The core was obtained in October 2008 with a modified Livingstone square-rod sampler. It is 2.85m long, spanning from 5.25m to 8.10m depth below the fen surface at Blacktail Pond. The core was wrapped in plastic and foil and has since been

analyzed and refrigerated in the MUS Paleoecology Lab. Age and depth markers were placed soon after collection to ensure accurate measurements as sediment cores dry and shrink. We took about a quarter of the core's diameter from each selected segment, aiming for about 2-3g of sediment/sample.

After samples were selected, each model was simplified utilizing only the mean age and depth of each sample. This also enabled us to obtain independent measurements for E/S (Equation 3) by measuring carbonate percentage via titration.

Sample Preparation. Samples were stored in individual plastic baggies before being dried at 90°C in an oven for about 2 days. They were then re-bagged. Samples were homogenized with a mortar and pestle in their individual baggies to avoid contamination. About 1g of each sample was set aside for meteoric ^{10}Be processing. Of the remaining sample, some was used in test procedures and carbonate digestions, while the rest was stored.

Laboratory Methods. Each 1g sample was digested via a series of acid additions. After each acid addition, sediment was allowed to react in an open vessel off of heat. Vessels were then placed on heat and allowed to react, before they were sealed and left on heat, usually for several hours. The samples were then dried down before the next acid addition. Hydrochloric acid digested carbonates. The first addition was 0.5M HCl, which was added, allowed to react, and dried down repeatedly until there was no visible reaction. The process was repeated with HCl of increasing molarity until there was no visible reaction upon addition of and heating in concentrated HCl. Then, aqua regia was

added with and increasing proportion of concentrated nitric acid, until only concentrated HNO_3 was used and there was no visible reaction. Then, nitric acid and hydrogen peroxide digested organics. These were small additions of $<1\text{mL}$ of each at a time, repeated until there was no visible reaction. Finally, mixtures of HNO_3 and hydrofluoric acid were added, with increasing additions of concentrated HF. Hydrofluoric acid removed the remaining silicates. Mixtures of each solvent were added until the sediment appeared to be dissolved.

After samples were digested, 1000mg/L Beryllium-9 SPEX carrier was added to each sample, as well as to two blanks before being run through cation and anion columns.

Data Analysis. The average blank measurements were subtracted from each sample measurement. Concentrations were plotted and compared to models and raw data from Blacktail Pond obtain by Krause and Whitlock, 2013. See main text for details and interpretation.

Supplement 3B: Estimating In-Situ vs. Terrigenous Sediment Inputs

In paleoclimate studies, carbonate production is often used as an indicator of overall lake productivity since the photosynthesis of macroscopic plants or algae consumes CO₂, which is thought to increase carbonate ion activity in the water. Carbonate precipitation is triggered by algal photosynthesis in the summer. The carbonate binds to available Ca²⁺ and precipitates as calcium carbonate (Leng and Marshall, 2004). Rates of photosynthesis are higher with warmer and sunnier conditions, so higher rates of calcite precipitation is thought to be associated with such conditions during summer months in a given setting (Meyers and Ishiwari, 1993). Since calcium enters the lake system via both biogenic and weathering and erosion processes, quantifying the ratio of Ca to Ti, provides an estimate of how much calcium was produced in the lake, versus calcium that was introduced to the lake from terrestrial sources because titanium is allochthonous and immobile. Titanium is a detrital sediment indicator because it is only produced allogenicly through the erosion of Ti bearing rocks (Cohen, 2003). Ti is immobile and insensitive to dissolution (Demory, et al, 2005), meaning that it can consistently act as a measure of terrigenous input to the lake. Since calcium enters the lake system via both biogenic and weathering and erosion processes, quantifying the ratio of Ca to Ti, provides an estimate of how much calcium was produced in the lake, versus calcium that was introduced to the lake from terrestrial sources.

Only relative abundances of Ca and Ti are needed to characterize the Ca:Ti record. In the Krause and Whitlock study, the Blacktail Pond core was run through an ITRAX XRF scanner at contiguous 0.5-cm intervals to quantify relative abundances of a

suite of elements. We utilized the Ti data from these measurements in our initial models as an estimate of the amount of terrestrial sediment in the core.

Supplement 3C: Comparisons of Measured Concentrations to known paleoenvironmental proxies

Here, we compare measured ^{10}Be concentrations to other raw data from Blacktail Pond in order to better understand the relationship of meteoric ^{10}Be to other paleoenvironmental proxies.

Indicators of Terrestrial Input. First, we compare measured concentrations with indicators of terrestrial input since ^{10}Be will be delivered to the lake via terrigenous material (See Supplementary Material 3A and 3B). These include XRF derived counts of titanium (binned by sample), percent carbonate and total carbon from Krause and Whitlock (2013), our own laboratory measurements of carbonate, and magnetic susceptibility from Krause and Whitlock 2013.

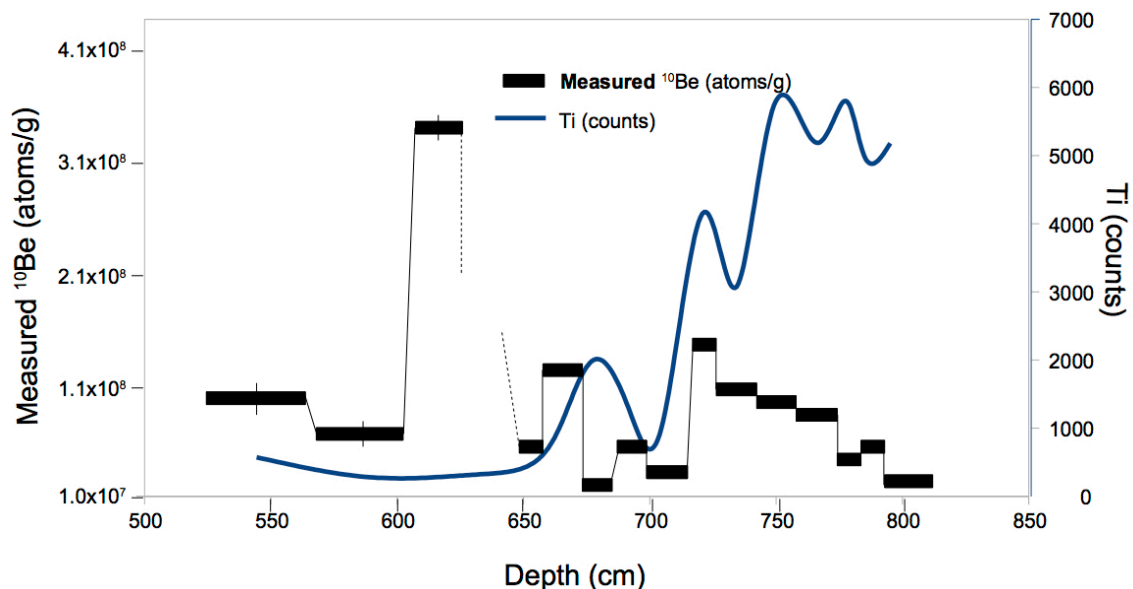


Figure 3.14 Measured ^{10}Be concentrations (black bars) on left y-axis and blue Ti counts on right y-axis. Ti counts are binned by our 16 samples from XRF data from Krause and Whitlock (2013).

There is little about relative abundances of Ti that matches patterns of ^{10}Be (Figure 3.14). At the beginning of the record, the relative abundance of titanium is high, while the ^{10}Be concentration is low. From there, the abundance of titanium drops until sample M9. Meanwhile, concentrations of ^{10}Be rise to sample M10 and then drop to M9. There are peaks at M10 and low points at M9 for both data sets. After this, there is little correlation between ^{10}Be and titanium (Figure 3.14).

Next, we compare measured concentrations to carbon-based measurements, including carbonate percentages measured in our lab (Figure 3.15a) and by Krause and Whitlock (2013) (Figure 3.15a), and total carbon measured by Krause and Whitlock (2013) (Figure 3.15a), and Ca:Ti ratios from Krause and Whitlock (2013) (Figure 3.15b).

The carbonate and total carbon weight percentages measured by Krause and Whitlock (2013) show little in common with ^{10}Be concentrations (Figure 3.15a). Strangely, the percent carbonate we measured shows some similarity to our data, with some offset (Figure 3.15a). In both data sets, ^{10}Be and carbonate tend to be high for the same intervals and low for the same intervals. This is contrary to expectations since carbon should be an indicator of lake-produced sediment.

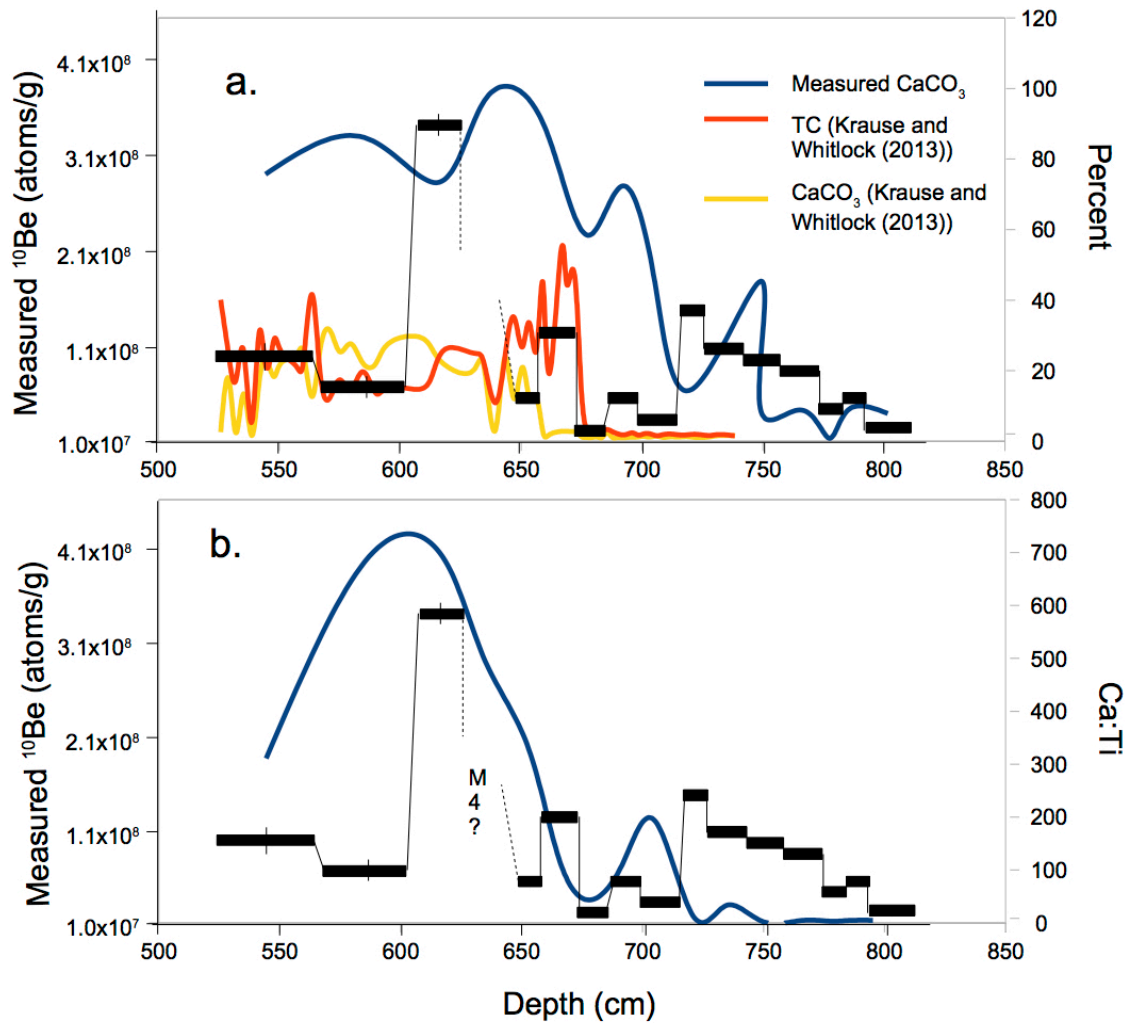


Figure 3.15. a. Measured ^{10}Be concentrations (black bars) on left y-axis and percentage total carbon or carbonate on the right y-axis. Our laboratory measurements of CaCO_3 are in blue (the upper most curve), CaCO_3 measurements from Krause and Whitlock (2013) is in yellow (lowest curve), and TC from Krasue and Whitlcok (2013) is in orange (middle curve). b. Measured ^{10}Be concentrations (black bars) with XRF derived Ca:Ti on right y-axis in blue. Differences between CaCO_3 percentages stem in part from lower sampling resolution and incomplete carbonate digestion in our lab.

Elemental Comparisons. Next, we compare relative abundances of each element measured by XRF to our measured ^{10}Be concentrations to add insight to the chemical behavior of ^{10}Be in a lake environment. Again, there is little correlation between ^{10}Be and most other elements. However, measured concentrations of ^{10}Be match more closely with percentages of CaCO_3 (Figure 3.15a) and counts of Ca, Sr, Se, Hg, As, and S (Figure 3.16 B, D, F) than any of our model predictions.

Other Data. We also compare measured ^{10}Be concentrations to pollen and charcoal counts from Krause and Whitlock (2013), as well as $\delta^{18}\text{O}$ levels. The pollen record for Blacktail Pond starts around our M9 sample, at 12.5ka and extends to the very top of the core, past the sampling intervals we used for meteoric ^{10}Be analysis. There is little correlation between pollen and ^{10}Be concentration (Figure 3.17).

The charcoal record for Blacktail Pond starts around 12.5ka (M9) and extends to 5ka, beyond our ^{10}Be record. Again, there is little correlation between ^{10}Be and charcoal, though charcoal has peaks near M3 and M6, two of our peaks (Figure 3.18). The $\delta^{18}\text{O}$ record starts at around 14ka (M10). Surprisingly, there is some correlation between the stable isotope record and the measured concentrations of ^{10}Be (Figure 3.12 in main text). Of all the raw data sets from Blacktail Pond, $\delta^{18}\text{O}$ seems to provide the closest match, in that there are low $\delta^{18}\text{O}$ values (more ^{16}O) where there is low ^{10}Be starting at sample M9, followed by gradual increases in both data sets. There are well-defined peaks at samples M6 and M3, corresponding to 11.1 and 9.6ka, respectively. $\delta^{18}\text{O}$ is high (more ^{18}O) in these intervals as well.

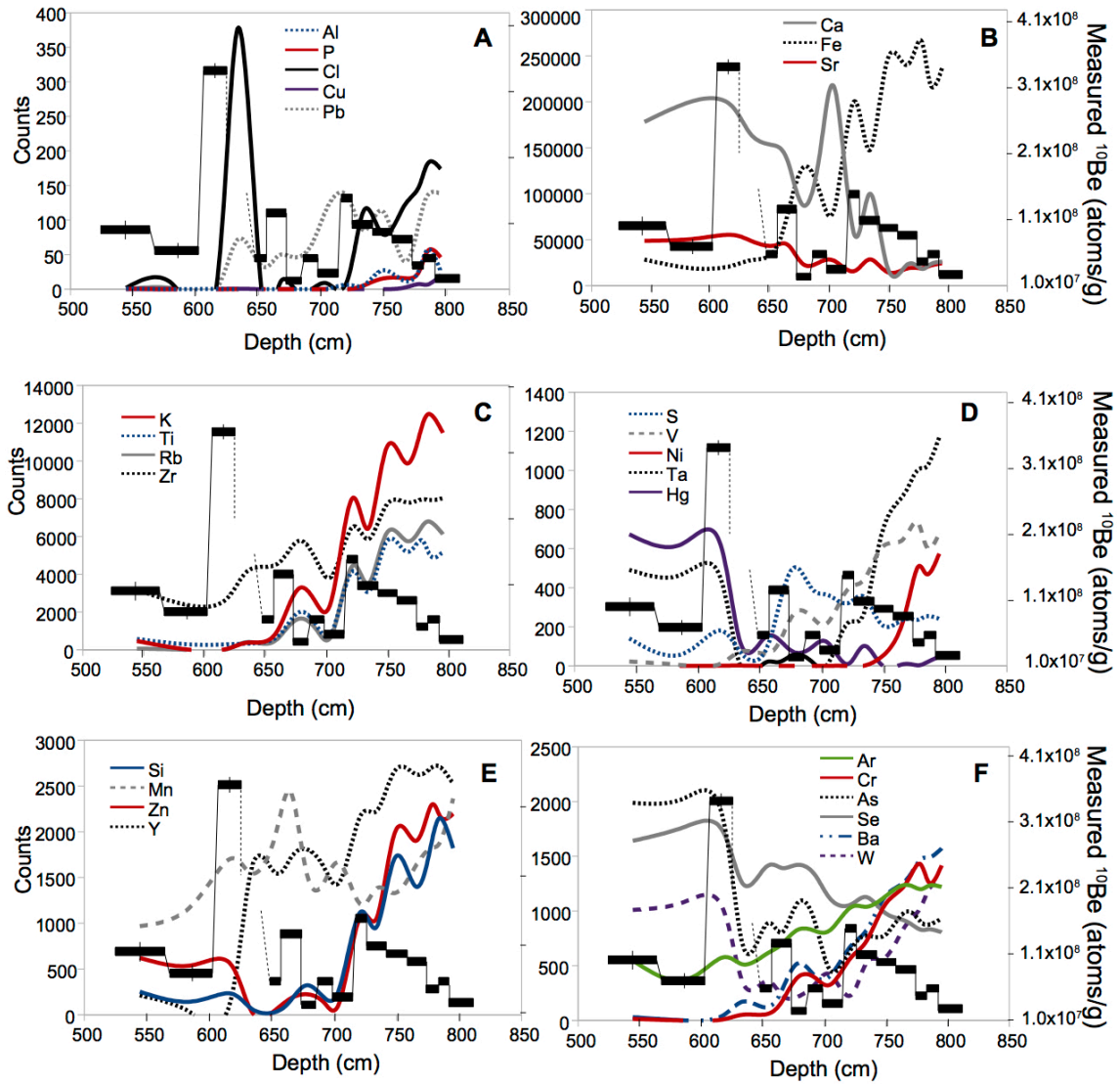


Figure 3.16 Comparisons of binned XRF derived elemental counts and our measured ^{10}Be . Of particular interest are Ca (B), Sr(B), Se(F), Hg(D), As, and S (D).

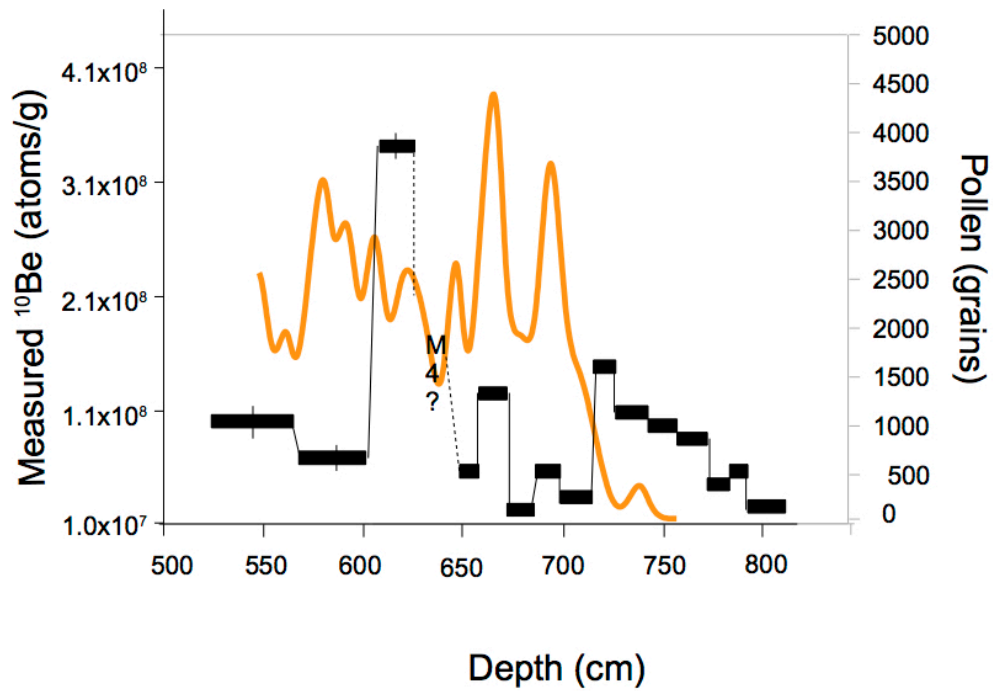


Figure 3.17 Total Pollen counts (orange) on right y-axis and measured ^{10}Be concentrations on left y-axis.

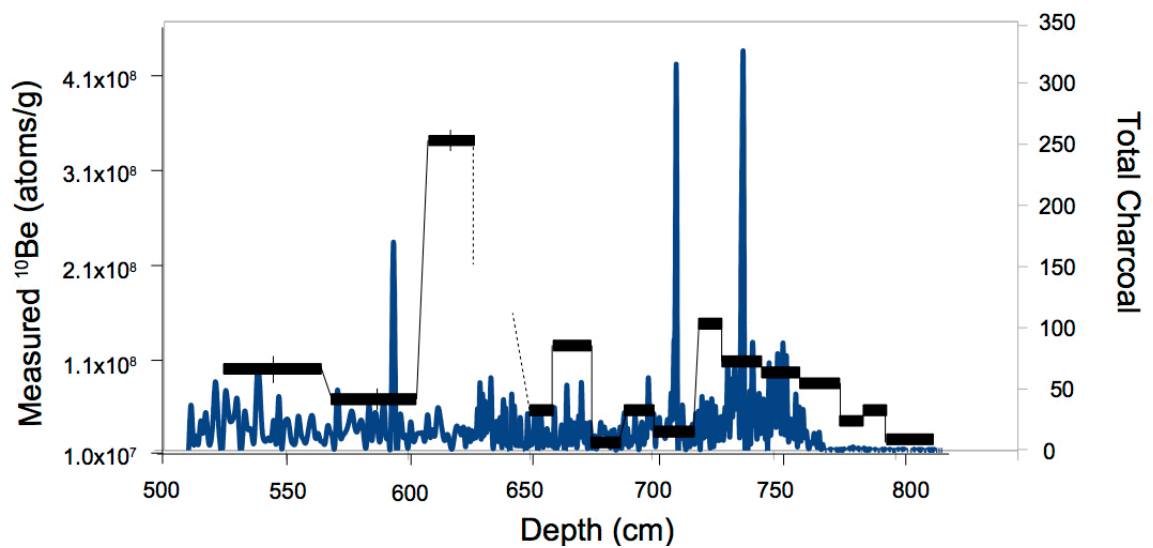


Figure 3.18 Total charcoal counts (blue) on right y-axis and measured ^{10}Be concentrations on left y-axis.

CHAPTER 4

CONCLUSIONS

Identifying and validating new tools that measure and trace Earth-surface processes is necessary to understanding the complex geomorphological, geochemical, and climatic controls that govern the many multifaceted relationships between climate and the landscape. Our understanding of surface processes, including weathering and erosion, and the influences that changes in climate, rock, soil, and biological processes have on these processes has been greatly deepened by the use of cosmogenic nuclides, especially beryllium-10. ^{10}Be has helped quantify the rates of the chemical and physical processes that transform soils and their sensitivity to forces such as climate. Much of this work has focused on “in-situ” ^{10}Be , because it is produced at known rates within mineral lattice structures on Earth’s surface (Dixon and Riebe, 2014). In contrast, this work, in conjunction with other recent works, shifts focus to meteoric ^{10}Be .

The characteristics of this underutilized nuclide, from its delivery in precipitation (Graly et al 2011; Graham et al. 2003) to its strong adsorption to sediments in a wide variety of landscapes (starting with Pavich et al., 1984; 1985; 1986 and later corroborated by many others), makes meteoric ^{10}Be an intriguing paleoenvironmental tool. Despite the promise of meteoric ^{10}Be in such studies, little published work has explored the use of meteoric ^{10}Be in lake archives, and interpretations of ^{10}Be concentrations in these accumulated sediments vary dramatically (Horiuchi et al., 1999; Horiuchi et al. 2001; Belmaker et al., 2008; Belmaker et al., 2014; Berggren et al., 2010; Nilsson et al.,

2011; Mann et al., 2012). By developing a model that predicts meteoric ^{10}Be concentrations with depth in accumulated sediments, I hope to provide a framework to aid in the interpretation of ^{10}Be records. Beryllium isotope concentrations in sediment may reflect atmospheric production of ^{10}Be and terrestrial and soil processes, including those that control lake sedimentation and sediment chemistry.

A theoretical and quantitative framework was developed to identify and assess the controls on the variability of ^{10}Be in accumulated lacustrine sediments (Chapter 2). This simple, quantitative model evaluates the controls on meteoric ^{10}Be in accumulated lake sediments. This model predicts the concentration of ^{10}Be as a function of atmospheric flux, sedimentation rate, and terrestrially delivered ^{10}Be (i.e. upland erosion or wind-blown dust). The model and the sensitivity of individual model parameters to changing conditions were tested in application to previously measured and published meteoric ^{10}Be profiles. These data sets come from lake systems in two distinct settings and latitudes, former Lake Lisan, Israel, and Lake Baikal, Russia.

Previous interpretations of ^{10}Be in the sediments of former Lake Lisan, Israel suggested that ^{10}Be concentrations reflect complex hydroclimatic mechanisms that impact sediment chemistry and dust delivery (Belmaker et al. 2008; Belmaker et al 2011; Belmaker et al. 2014). ^{10}Be concentrations in Lake Baikal, Russia are attributed to changes in the delivery of terrestrial material during glacial periods (Horiuchi et al. 1999; Horiuchi et al. 2001). Quantitative analysis of both Lake Lisan and Lake Baikal suggest that sedimentation rate is the primary control on temporal patterns of ^{10}Be concentrations (Figures 2.4 and 2.6). Sedimentation rates themselves may also reflect local climate. In

both lakes, atmospheric flux alone did not explain the total magnitude of ^{10}Be concentrations in a core (Figures 2.5 and 2.7). To achieve the measured concentrations from these studies the model indicates that ^{10}Be had to have been delivered to the lake from upland detrital sources through surficial transport or wind.

Although ^{10}Be in accumulated sediments did not provide much direct insight to changes in precipitation, it did show dilution by lake sedimentation and addition of further ^{10}Be from upland erosion are critical in considering ^{10}Be concentrations in lake sediments. Both of these factors are indirectly related to changes in hydroclimate.

We show that while a variety of environmental conditions influence ^{10}Be in lake sediments, these can be quantified with surprisingly simple parameters. The model analyzed the sensitivity of meteoric ^{10}Be to environmental changes including past hydroclimate, erosion regimes, and dust deposition based on only sedimentation rate, density, and terrigenous sediment fraction. This new quantitative framework suggests that ^{10}Be concentrations measured in lake sediments can be used to assess changes in sedimentation rates, inputs of dust or upland sediments, and hydroclimate, though it does not provide a direct proxy for precipitation. The changes in model parameters required to match measured concentrations highlight the potential to identify environmental changes that influence each of these factors.

Next, ^{10}Be concentrations were measured in the sediments of Blacktail Pond, Yellowstone National Park. Blacktail Pond provided a new setting for evaluating the model framework (Chapter 2), because it appears to be a simple system in which the only additions of ^{10}Be to the lake are from the atmosphere and from terrigenous material.

Additionally, Blacktail Pond provided the opportunity to compare ^{10}Be concentrations with independent paleoenvironmental proxies, including oxygen isotope, pollen, and charcoal records, as well as variation in geochemical composition of the sediments.

The model framework was applied to this system, predicting ^{10}Be concentrations in the Blacktail Pond core, relying on paleoclimate data from the region to define model parameters (Bartlein et al. 1998; Krause et al., 2015; Whitlock et al., 2012; Whitlock and Bartlein, 1993; Millspaugh et al., 2004, Licciardi et al. 2004 and others). These predictions informed sample selection from the Blacktail Pond core. The measured concentrations of meteoric ^{10}Be in the Blacktail Pond record did not resemble any modeled predictions. Contrary to the model expectations, ^{10}Be concentrations were not diluted by sedimentation rate.

Required changes in the ratio of catchment area to lake area may explain the fluctuations in ^{10}Be concentration at Blacktail Pond.. When lake size is accounted for, high levels of ^{10}Be correspond to a larger catchment area relative to lake area, so assuming the catchment area does not change, a smaller lake area is needed to achieve higher ^{10}Be concentrations and vice-versa. Because high ^{10}Be values correspond to high $\delta^{18}\text{O}$ values, contribution of ^{10}Be from runoff is a possibility. To confirm this idea, however, more work involving comparisons of ^{10}Be and $\delta^{18}\text{O}$ records of both open and closed-basin lakes is needed. Even without a relationship to $\delta^{18}\text{O}$ records, catchment area in studies of ^{10}Be concentrations in lake sediments should be considered to ensure accuracy of modeled concentrations.

Concentrations of ^{10}Be in Blacktail Pond may also be explained by a soil or lake environment process we have not yet considered. Most measured concentrations of ^{10}Be may be accounted for in our model by changing the concentration of ^{10}Be on incoming terrigenous sediment, a product of both climate and landscape processes, though it is hard to justify every possible change. As in any natural system, the processes that ultimately influence ^{10}Be concentrations in the sediments of Blacktail Pond are complex. ^{10}Be concentrations in this system may be especially sensitive to even small changes in model parameters. Clearly, more work needs to be dedicated to understanding why ^{10}Be concentrations in a variety of lake settings might deviate from those predicted by our model.

Finally, laboratory error cannot be ruled out. While previous blank measurements indicated that the lab was suitable for meteoric ^{10}Be processing, every step of processing from digestions to target packing, allowed opportunity for both sample loss and contamination. Great care should be taken by researchers to establish a standard laboratory procedure for extracting meteoric ^{10}Be from lacustrine sediments. Future work in the area should reevaluate Blacktail Pond ^{10}Be concentrations and should seek to establish a ^{10}Be record for the region by comparing ^{10}Be records from several lakes, taking into account lake size, landscape stability, atmospheric flux and sedimentation rate.

Final Remarks

This work advances a new tool for environmental tool for environmental reconstruction and opens new lines of inquiry regarding the behavior of meteoric ^{10}Be in

a lake environment. It is the first work to propose a quantitative model framework that explains ^{10}Be concentration in accumulated lake sediments. This framework provides a basis for interpretation of ^{10}Be records in accumulated sediments and their sensitivity to our model parameters. ^{10}Be in paleorecords provides only indirect insight into past precipitation, but the model shows that atmospheric flux is not the only influence on ^{10}Be concentration in accumulated sediments. Parameters such as sedimentation rate, the percentage of terrigenous material, and catchment area should not be ignored. Overlooking these parameters, especially sedimentation rate, may lead to the misinterpretation of ^{10}Be records. Future work should continue to test this model in a wider variety of latitudes and in areas where dramatic changes in precipitation may have influenced the atmospheric delivery of ^{10}Be , especially on long timescales.

While this work has made it clear that much further investigation is required to understand the behavior of and controls on meteoric ^{10}Be in lake sediments, it also highlights the ability to model reasonable concentrations given the right parameters. Meteoric ^{10}Be in lake sediments is underutilized and, hence remains an intriguing possible paleoenvironmental proxy. More research should be dedicated to measurements of this nuclide in a variety of lake records.

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APPENDICES

APPENDIX A

LITERATURE REVIEW

Meteoric ^{10}Be

The Earth is constantly bombarded by high-energy particles, called cosmic rays, produced by the shockwaves of supernovae. These particles consist of atomic nuclei and individual subatomic particles, including protons, neutrons, electrons, positrons, pions, muons, and alpha particles. In the upper atmosphere, cosmic rays are primarily composed of protons and alpha particles. Secondary particles, like neutrons and muons, are produced by the initial reactions of cosmic rays higher in the atmosphere in a cascade of spallation reactions (Figure 1.1) (Masarik and Beer, 1999; Masarik and Beer, 2009). ^{10}Be and other cosmogenic nuclides, such as ^3He , ^{14}C , ^{21}Ne , ^{22}Ne , ^{26}Al , ^{36}Cl , and ^{53}Mn (Dunai, 2010), are the products of interactions of primary and secondary cosmic ray particles with atomic nuclei in the upper atmosphere or at Earth's surface. We focus on ^{10}Be , usually produced through the spallation of oxygen and nitrogen by these energetic particles. Nucleons, protons in the upper atmosphere and neutrons at the surface, dominate cosmogenic nuclide production above the surface, but in the subsurface, production is muon dominated. This is because muons interact weakly with matter, and hence, maintain energy that allows them to reach greater penetration depths than neutrons, which lose energy with each interaction (Lal, 1988; Dunai, 2010). ^{10}Be is not present on Earth's surface or in the atmosphere unless it has been exposed to cosmic radiation (Lal and Peters, 1967; Lal, 1991). Two varieties of ^{10}Be are produced by different mechanisms: meteoric ^{10}Be is produced in the upper atmosphere, while the production of in-situ produced ^{10}Be occurs at known rates on Earth's surface when oxygen atoms

within the lattices of mineral grains are bombarded with cosmic rays (Lal and Peters, 1967).

Both nuclides accumulate on Earth's surfaces predictably over time such that their concentrations reflect the amount of time a surface has been exposed to cosmic radiation (Willenbring and von Blanckenburg, 2010). Generally, this means that surfaces with higher concentrations of nuclides have been exposed to cosmic radiation longer and are, therefore, older or more stable than surfaces with lower concentrations (Dixon and Riebe, 2014). ^{10}Be concentrations in a range of geologic materials have been used to calculate landscape age, rates of erosion and weathering as well as to record sediment burial and track soil movement (Willenbring and von Blanckenburg, 2014).

Meteoric ^{10}Be was developed as a tool to date soil age and residence time in the late 1980s and early 1990s, as well as to quantify erosion and trace other hillslope processes (Pavich et al., 1984; 1985; 1986). However, the limited understanding of meteoric ^{10}Be delivery to the surface and its chemical behavior on the surface at the time limited its widespread use, and its popularity was soon eclipsed by the rise of the use of in-situ ^{10}Be on the grounds that its production rates are well-quantified (Lal, 1991). Because in-situ ^{10}Be can be used to quantify many of the same processes and because its production rates were more well constrained, meteoric ^{10}Be was soon set aside by many researchers in favor of in-situ ^{10}Be (Dixon and Riebe, 2014).

Recently, however, progress has been made in understanding the controls on the production of ^{10}Be in the atmosphere and its delivery to the surface, as well as its chemical behavior in soils and sediments, revealing its potential as a versatile

geochemical tracer in a wide variety of environments. Since quartz need not be present in an environment to utilize meteoric ^{10}Be , it has also been shown to be a powerful tool for tracing fluvial sediment sources by comparing concentrations in stream sediment yields to concentrations in source areas (Reusser et al. 2010) and estimating catchment wide erosion rates in a variety of geographic and climatic settings (Willenbring and von Blanckenburg, 2010). Furthermore, its delivery to the surface has been shown to vary strongly with precipitation (Graly et al., 2011), and combined with its strong adherence to fine soils and sediments, ^{10}Be shows promise as a novel paleoenvironmental proxy.

Controls on the Production of Meteoric ^{10}Be

Meteoritic ^{10}Be is produced via the spallation of oxygen and nitrogen by the high-energy particles known as cosmic rays (discussed above) (Masarik and Beer, 1999; Masarik and Beer, 2009). The production of ^{10}Be is controlled by the intensity of cosmic ray flux, which is modulated by changes in solar activity and the orientation and intensity of Earth's geomagnetic field. These controls limit the kinds of cosmic particles that interact with the atmosphere and produces a latitudinal pattern of production rate, in which more radiation reaches the poles than mid-latitudes (Figure A1) (Lal and Peters, 1967; Masarik and Beer, 1999); hence production rates tend to be higher at high latitudes and lower in the tropics. Earth's geomagnetic field acts as a shield, deflecting incoming particles based on charge, energy, and angle of incidence (Lal and Peters, 1967; Masarik and Beer, 1999; Masarik and Beer, 2009), preventing most low-energy particles from interacting with the atmosphere (Masarik and Beer, 1999). Nuclide production is higher in the stratosphere than the troposphere (Lal and Peters, 1967). In the case of in- situ

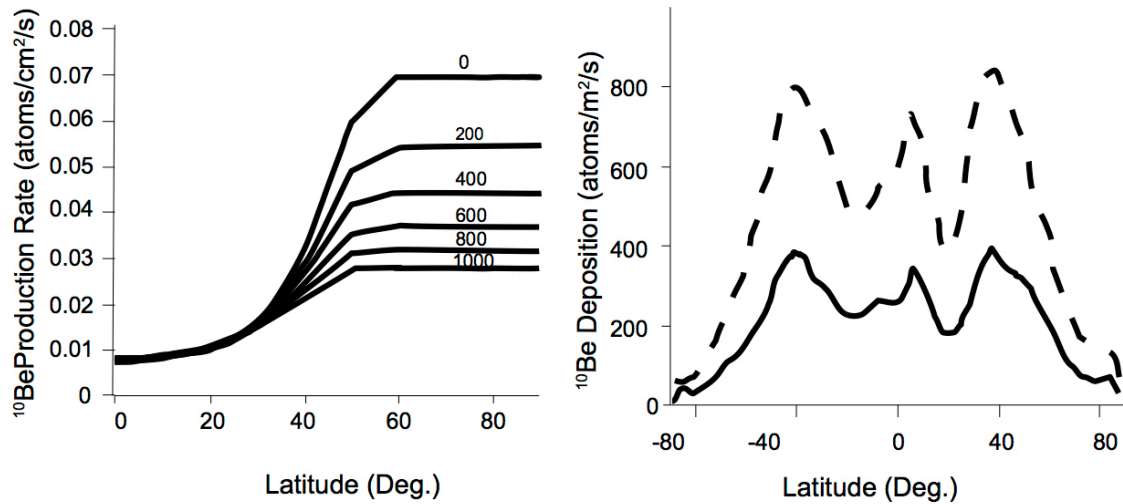


Figure A1 Global patterns of ^{10}Be production and delivery. A) Latitudinal variation in atmospheric production rate of meteoric ^{10}Be with different levels of solar activity (0-1000 MeV) (adapted from Masarik and Beer, 1999). B) Total (wet and dry) deposition of ^{10}Be with latitude, considering a model with the parameters of the Laschamp geomagnetic excursion (dotted line) and a model with average modern parameters (Adapted from Heikkilä, et. al., 2009). Patterns of deposition partially match patterns of deposition, with significant modulation by precipitation rates, such that patterns of precipitation with latitude match patterns of ^{10}Be deposition with latitude.

produced ^{10}Be , altitude must be taken into account when calculating production rate, because particles lose energy as they travel towards the surface, so production rates at sea level are lower than at higher altitudes (Lal and Peters, 1967). In contrast, most of the production (~99%) of the meteoric ^{10}Be takes place above 3km in elevation before it is delivered to the surface and altitudinal scaling is not needed to account for its production rates (Willenbring and von Blanckenburg, 2010).

Changes in geomagnetic field strength have a stronger influence on mid- to low-latitudes than high latitudes (Masarik and Beer, 1999). Conversely, solar modulation of ^{10}Be production is strongest at high latitudes and altitudes (~ 60 - 80 degrees) and

changes in solar activity mostly alter the production of ^{10}Be at high latitudes, while having very little effect on production rates near the equator (Figure A1) (Masarik and Beer, 1999). Solar modulation controls cosmic ray variability, increasing particle flux energy (Masarik and Beer, 1999). Research of accumulated concentrations of meteoric ^{10}Be has primarily focused on high-latitude ice cores in an attempt to reconstruct these controls on ^{10}Be production (i.e. Finkel and Nishiizumi, 1997; Muscheler et al, 2004, Vonmoos et al., 2006), although a few high-latitude studies examine sediment cores (i.e. Berggren et al., 2010 and others). While high-resolution ice core records show short-term variation in ^{10}Be due to solar variability, long-term flux does not seem affected (Vonmoos et al. 2006). Additionally, interpretation of ^{10}Be concentrations can be difficult in ice records because climate variations and dust deposition may mask the production signal (Wagner et al., 2000 and 2001; Lal, 2007).

Controls on the Delivery of Meteoric ^{10}Be

The flux of meteoric ^{10}Be to the surface varies by location and with time based on changes in production rate and by different delivery mechanisms of the nuclide from the atmosphere to the surface. After production in the atmosphere, meteoric ^{10}Be forms highly reactive molecules like $^{10}\text{BeO}_2$ or $^{10}\text{Be}(\text{OH})_2$, attaches to aerosols, and is distributed via the movement air masses (Masarik and Beer 1999). It is primarily produced in the stratosphere where it is strongly mixed before deposition, usually with a residence time of 1-2 years (Masarik and Beer, 1999; Raisbeck et al., 1981b), such that the highest fluxes occur at mid-latitudes and there is good agreement between ^{10}Be records in the northern and southern polar regions (Heikkilä et. al., 2009). ^{10}Be is then

deposited via both dry and wet processes (Field et al., 2006). Mixing is thought to be a significant factor in controlling deposition patterns. Older models for the production rates of ^{10}Be assumed homogenous atmospheric mixing such that deposition was proportional to the production rate averaged over the corresponding region (McCracken 2004; Usoskin, 2004). Since then, models have used a general atmospheric circulation model for ^{10}Be (Field et al., 2006; Heikkilä et al., 2009; Gennady et al., 2010). This means that records of ^{10}Be at any latitude should reflect changes in global mean production rate and therefore, act as a proxy for past solar activity (Heikkilä et al., 2011). There are, however, differences in records from different regions, especially on short timescales, complicating the use of ^{10}Be as a proxy for cosmic ray intensity. It is generally accepted these variations are due to climate. For example, high-latitude climate changes might be imprinted on meteoric ^{10}Be concentration in ice cores if precipitation is delivered primarily as snow, sometimes obscuring changes in concentration that depend on production rate (Heikkilä et al., 2009).

Differences in the prominence of varying deposition mechanisms may cause concentrations to vary by location. In most mid- and low-latitude locations, delivery rate of meteoric ^{10}Be seems to vary strongly with precipitation (Figure 2.1a) (Graly et al., 2011; Graham et al., 2003). This idea was asserted early on as a result of the scavenging of ^{10}Be from the atmosphere by precipitation and how it might cause the ^{10}Be signal in a given record to deviate from the signal predicted by production rate (Lal, 1987). While atmospheric mixing should smooth ^{10}Be latitudinal production so that it reflects global mean production rate, rather than the production rate for that latitude, the differences in

^{10}Be at high latitudes are often attributed to uncertainty. At lower latitudes, they may reflect changes in the fraction of wet to dry deposition of ^{10}Be due to the scavenging of ^{10}Be by precipitation, possibly reflecting local climatic conditions (Heikkilä et al., 2013). However, if deposition fluxes are normalized for precipitation, patterns of deposition roughly match patterns of production (Figure A1) (Graly et al., 2011). Furthermore, models that map global ^{10}Be deposition show that ^{10}Be deposition is high in regions where precipitation is high. Yet, in areas with low precipitation, there is still significant dry deposition of ^{10}Be which depends mostly on production rates and aerosol mass as not all of the ^{10}Be will be scavenged from the atmosphere in arid environments (Heikkilä et al. 2013; Weng et al. 2015). In most environments, dry fallout contributes a maximum of about 8-10% to the total fallout (Wallbrink and Murray, 1994; Feely et al., 1989). While there is little data that correlates precipitation with ^{10}Be concentrations in terrestrial systems, past work has shown that meteoric ^{10}Be deposition, when corrected for dust deposited ^{10}Be , can be predicted by latitude and average annual precipitation (Figure 2.1b) (Graly et al. 2011; Graham et al., 2003). There is, however, uncertainty in predicting long-term ^{10}Be flux to the surface based on precipitation in any specific locations due to the contribution of dust to ^{10}Be inventories (Lal, 2007; Graly et al., 2011) and the variability of ^{10}Be concentration in individual rainfall events (Brown et al. 1989).

It is important to note that there is no consensus concerning the relationship of ^{10}Be to precipitation. While there is good evidence in many settings for an additive effect as explained above, there is also significant evidence of a dilution effect. Some settings show that periods of less precipitation result in higher concentrations of ^{10}Be , implying

flux to the surface remains the same (Yiou et al., 1997). There is even evidence suggesting that in regions of very high rainfall (more than 6 cm/month), ^{10}Be may be completely scavenged from the atmosphere, such that increasing rain dilutes concentrations further (Figure 2.1b) (Monaghan et al., 1985). It has been suggested that ^{10}Be delivery in some environments is diluted by precipitation, especially in settings with high volumes of precipitation in which all the available ^{10}Be is scavenged from the atmosphere (Yiou et al., 1997). Importantly, this dilution is noted in concentrations of ^{10}Be in precipitation (Monaghan et al., 1985; Field et al., 2006; Willenbring and von Blankenburg, 2010). However, these concentrations vary over a relatively small range compared to the range of precipitations values, thus ^{10}Be flux is still largely controlled by rainfall rate regardless of this dilution effect. Additionally, if an average or constant ^{10}Be deposition rate is accounted for, precipitation should deliver more ^{10}Be to a given location until all of the ^{10}Be is scavenged from the atmosphere, at which point, the ^{10}Be concentration in accumulated sediments as a result of a given rainfall event would plateau.

Behavior of Meteoric ^{10}Be in a Terrestrial Setting

The concentration of meteoric ^{10}Be represents a combination of the interplay between atmospheric production rate, delivery and behavior on the surface. Once delivered to the land surface as particle reactive $^{10}\text{BeO}_2$ or $^{10}\text{Be}(\text{OH})_2$, ^{10}Be binds readily to fine soil and sediment (Graly, et al., 2010; Pavich et al., 1984). ^{10}Be has a high partition coefficient of around 10^5 mL water/g soil at pH 6 and above, depending on environment, indicating it is relatively immobile (You et al., 1989; Shen et al., 2004; Graly et al., 2010). This allows meteoric ^{10}Be to be retained on soil particles or

incorporated as a cation into clays, organic matter, or oxyhydroxides under most natural chemical and mineralogical boundary conditions. This means that the pH of the soil should be greater than pH 6 (or at least between pH 3-11) so that particle reactive beryllium hydroxides and organic complexes are not lost to solution or competing with aluminum, which is mobilized at pH 6, and other metals for adsorption sites on clay particles. (Barg et al., 1997; Willenbring and von Blanckenburg, 2010).

Concentrations of meteoric ^{10}Be in soil also seem to decrease with increasing grain size (Figure A2), something that may also influence ^{10}Be distribution in the soil. Horizons that are clay rich or have smaller grain sizes have more surface area, allowing for more adsorption and ion exchange sites with which ^{10}Be can interact (Barg et al., 1997; Shen et al., 2004; Wittmann et al., 2012). The hydrology of the soil and landscape also play a role in retention. Dissolved ^{10}Be will only be lost to groundwater if the soils in a given environment are exposed to sufficient water for sufficient mobilization. This also contributes to the retention of ^{10}Be at soil surfaces (Barg et al., 1997; Shen et al., 2004).

Little is yet known about ^{10}Be redistribution in the soil column, other than the fact that it is hampered by high partition coefficients and influenced by soil porosity (You et al. 1989; Aldahan et al. 1999; Brown et al. 1992). Meteoric ^{10}Be accumulates with soil residence time and strongly traces particle movement (Brown et al. 1992), and general distribution patterns have been observed in soils in a variety of locations (Blake et al. 2002; Balco, 2004). ^{10}Be concentrations tend to decrease exponentially with depth

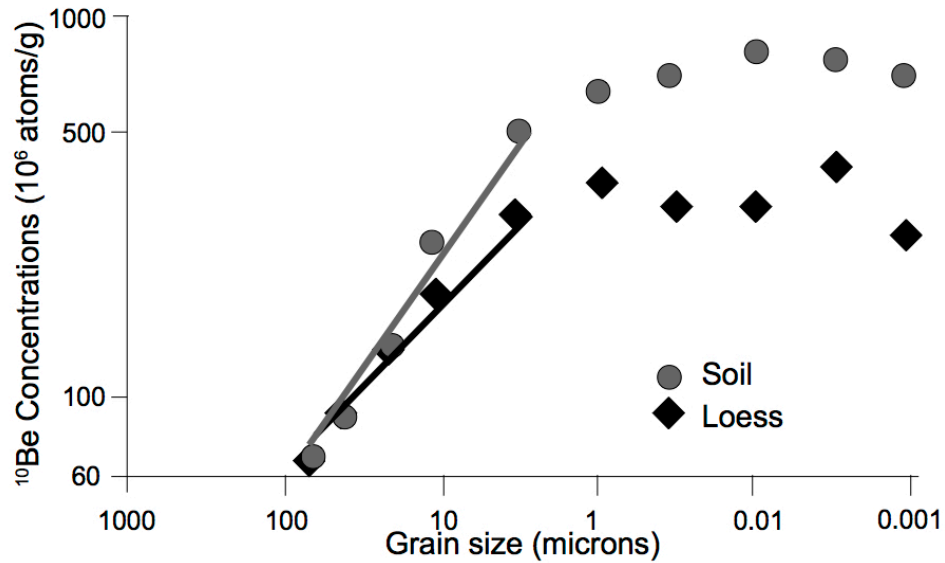


Figure A2. Grain size dependency of ^{10}Be in soil (circles) and loess deposits (diamonds),. (Adapted from Shen, 2004).

(Figure A3), especially in younger soils, unless there is an especially clay rich horizon (providing an abundance of adsorption sites) further down within a soil profile (Brown et al., 1992; Blake et al. 2002; Balco, 2004) Willenbring and von Blanckenburg, 2010). Generally, in young or quickly eroding soils, highest meteoric ^{10}Be concentrations are found near the surface, while in older or more slowly eroding soils, the highest meteoric ^{10}Be are found at depth, a pattern that correlates with total ^{10}Be inventory (Graly et al., 2010).

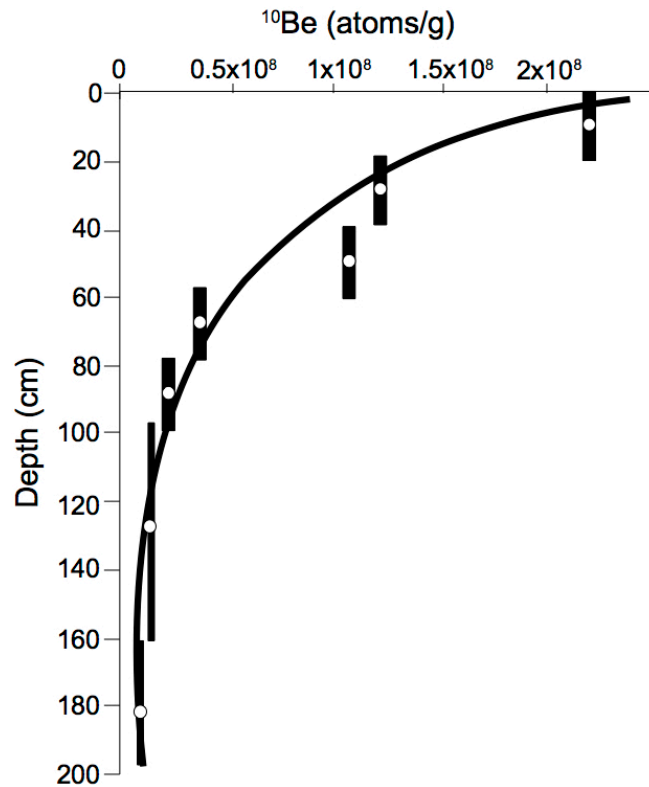


Figure A3. A depth profile of meteoric ^{10}Be from a Minnesota soil. Note the exponential decrease with depth, as shown by the black line (exponential best fit) which can be used to calculate mean soil residence time (Adapted from Balco, 2004).

Applications of Meteoric ^{10}Be to Studies of Landscape Evolution

Soil Systems. Early studies utilizing meteoric ^{10}Be in landscapes compared measured inventories, as calculated from concentration with depth, with calculated inventories based on delivery rate and decay rates (^{10}Be has a half-life of 1.39 million years) to determine erosion rates and soil residence times (Pavich et al. 1985; 1986). Some studies applied ^{10}Be movement on hillslopes to measure soil transport processes and weathering rates, described as bedrock-to-soil conversion rates (Monaghan et al.

1992). These studies often assumed a constant deposition rate to account for limited understanding of delivery.

As production rates and delivery mechanisms and chemical behavior of meteoric ^{10}Be become better constrained, the use of meteoric ^{10}Be is experiencing a resurgence. Recent studies reclaim some of the earlier uses for the nuclide. For example, West et al., (2013) measured meteoric ^{10}Be with to calculate mobile regolith flux along hillslopes, and Egli et al., (2010) used meteoric ^{10}Be inventories and decay rates to calculate soil age. Erosion rates can be calculated from similar measurements, assuming a steady state between ^{10}Be removal by erosion and flux to the surface (Willenbring and von Blanckenburg, 2010). However, as we gain further understanding of delivery rates and chemical behavior of ^{10}Be , further applications are being developed. For example, the isotope ratio of meteoric ^{10}Be to ^9Be (Appendix C), a stable isotope found in many minerals, can yield denudation rates and weathering intensities through mass balance analysis in the weathering zone. The method can be applied to an entire river basin when the flux of dissolved ^9Be and ^{10}Be collected from rivers is compared to the ^9Be contribution to the watershed system from weathering and the ^{10}Be contribution from precipitation (von Blanckenburg et al., 2012; Wittmann et al., 2012). Similarly, meteoric ^{10}Be retention can be approximated using loss of ^9Be to determine soil residence time more accurately than measurements of ^{10}Be alone (Bacon, et al. 2012) since it can be used to normalize differences in ^{10}Be concentrations with particle size (Willenbring and von Blanckenburg, 2010).

Since quartz need not be present in an environment to utilize meteoric ^{10}Be , it has also been shown to be a powerful tool for tracing fluvial sediment sources by comparing concentrations in stream sediment yields to concentrations in source areas (Reusser, et al. 2010) and estimating catchment wide erosion rates in a variety of geographic and climatic settings (Willenbring and von Blanckenburg, 2010). This later application, similar to in-situ methods, is possible using the measured concentration of meteoric ^{10}Be at the surface of the soil, calculating concentration at depth, and using spatially averaged erosion rates in river sediments (Willenbring and von Blanckenburg, 2010). Calculation of erosion rates or age using meteoric ^{10}Be usually requires the assumption that landscapes are in steady state. This means that the flux of meteoric ^{10}Be fallout is assumed to roughly be equal to the flux of ^{10}Be adsorbed to sediment particles that leave the system. This is most likely to be true in environments where infiltration is higher than runoff and soil pH is conducive to retention ($\text{pH} > 6$). In environments where this is not the case, the fact that accumulation rates of meteoric ^{10}Be on the surface are independent from distribution with depth indicates it could be used as a tool in a variety of settings (Willenbring and von Blanckenburg, 2010).

Sediment Archives. Studies of accumulated meteoric ^{10}Be often focus on ice cores in order to reconstruct controls on atmospheric production rates, but the study of ^{10}Be in accumulated sediments may shed light on how sediment transport processes and the deposition, both dry and wet, of ^{10}Be has changed with time. However, when analyzing ^{10}Be concentrations, these archives present a suite of challenges, some unique to each system and others common to all ^{10}Be studies. A variety of processes will exert

influence on ^{10}Be concentrations in accumulated sediments. Not only will these sediments be subject to terrestrial transport and soil processes, but in a lake environment, processes such as sedimentation rate and lake and sediment chemistry add even further complexity to understanding ^{10}Be concentrations in accumulated sediments.

The first measurements of ^{10}Be in lake sediments were obtained by Raisbeck, et al., 1981a. The primary observation from this study was that ^{10}Be delivered by erosion of the drainage basin was needed to account for concentrations of ^{10}Be in the sediment, and idea corroborated by Lundberg, et al., 1983. While some studies have reported ^{10}Be concentrations in past and present lake sediments more recently (Horiuchi, et al., 1999, Horiuchi, et al. 2001, Belmaker, et al., 2008, Belmaker, et al., 2014, Berggren et al., 2010, Nilsson et al., 2011 and Mann et al., 2012), the controls on ^{10}Be in accumulated lacustrine sediments are not well quantified. Dating methods of lacustrine sediment cores allow for higher resolution records than for ice core records, however lake systems are more complex, especially given the incomplete understanding of meteoric ^{10}Be in terrestrial systems. Interpretations of ^{10}Be concentrations vary widely and do not provide a quantitative framework for the suggested mechanisms. However, most studies of ^{10}Be in lacustrine sediments attribute concentrations to fluctuations in production rates (Berggren, et al., 2010; Nilsson, et al., 2011; Mann, et al 2012), though, as with ice cores, the production signal may be complicated by hydrologic and climatic conditions in the lake catchment (Horiuchi, et al., 1999, Horiuchi, et al. 2001, Belmaker, et al., 2008, and Belmaker, et al., 2014). Developing a model that predicts ^{10}Be concentrations with depth

in accumulated sediments provides a framework with which to interpret meteoric ^{10}Be records in different settings.

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APPENDIX B

BLACKTAIL POND

The Paleoenvironmental Record of Blacktail Pond

Yellowstone National Park boasts a complex geologic and ecological history. Blacktail Pond, a small lake in Northern Yellowstone provides a record of geological and environmental change since the end of the last glacial (Huerta et al., 2009; Krause and Whitlock, 2013; Pierce et al., 2014). It is located in a kettled glacial outwash terrace that transitions to a remnant meltwater channel that formed during ice retreat around 15,000 years ago when Blacktail Deer Creek was diverted from its course to the Yellowstone River (Pierce, 1979). Pollen and charcoal records from sediment cores have been the focus of three paleoecological studies (Gennett and Baker, 1986; Huerta et al., 2009; Krause and Whitlock, 2013). These studies allow for the comparison of ^{10}Be concentrations with a variety of independent paleoclimate proxy. Oxygen isotope, and geochemical records at Blacktail Pond have also been analyzed (Krasue et al. 2015), and paleoclimate models have been used to understand the large-scale patterns of climate change over the last 21,000 years (Bartlein et al, 1998; Alder and Hosteler, 2014).

Modern Setting

Today, seasonal patterns of precipitation in Yellowstone are driven by two large-scale precipitation regimes described as “summer-wet” and “summer-dry” (Whitlock and Bartlein, 1993). The specific precipitation regime is determined by topography. The summer-dry regime dominates in central and southern Yellowstone and is influenced by the features that control the Northwestern U.S., including mountainous areas of Montana and northern Idaho. Dry summers are the result of the influence of northeastern Pacific

subtropical high-pressure system. Subsidence over the ocean suppresses precipitation, generating dryness that flows east following the Columbia and Snake Rivers through Rawlins Gap, WY and the lowlands of the U.S./Mexico border. Winter precipitation under this regime results from storms tracking east along the jet stream from the Pacific Ocean, which shifts further south in winter, generating precipitation over the Yellowstone Plateau and the Beartooth Uplift (Whitlock and Bartlein, 2004). The summer-wet regime dominates in northern Yellowstone and is characterized by the features that control the Southwestern U.S., including the basins of Wyoming, Montana, and the Southern mountains of Idaho. Summer precipitation is produced by onshore flow of moist air from the Gulf of California and Gulf of Mexico due to the warming of the interior of the continent in the summer and summer monsoonal flow from the southwest is direct north through areas of the Great Basin until it reaches southeastern Oregon, southern Idaho, and southwestern Montana. Winters are dry as precipitation is diminished by the Cordilleran rain shadow, as western mountain ranges block Pacific moisture from reaching the continental interior (Whitlock and Bartlein, 1993). As spatial scale decreases, topography and local nonclimatic factors, like as substrate and disturbance regimes become more important in determining environmental response indicators (Mock and Bartlein, 1995; Whitlock and Bartlein, 1993). Today, northern Yellowstone continues to dry and supports steppe and montane conifer communities, while central and southern Yellowstone may be getting wetter and supports Englemann spruce and subalpine forest (Bartlein et al., 1998).

Quaternary History

The paleoenvironmental history of the GYE combines the geologic and glacial legacy of the landscape with changes in climate, from large-scale climatic shifts to more regional and local features. The late glacial-early Holocene transition (20-8ka) (Alley et al., 2002) featured rapid environmental change across the globe. As summer insolation rose and greenhouse gas levels increased, temperatures rose and ice sheets and glaciers retreated. This dramatic shift in climatic physical and climatic conditions led to complex changes in the landscape and the environment. The Blacktail Pond record covers ~14.6-6.9 ka, spanning this pronounced transition (Krause and Whitlock 2013).

Glacial History

Glaciers shaped much of the surface topography of the Rocky Mountain region. In the late Pleistocene, a large ice cap covered the Yellowstone Plateau, with outlet glaciers to the north, west, and south (Licciardi and Pierce, 2008). While there are areas of Southwest Montana and Northwest Wyoming, including within Yellowstone that display glacial deposits of Bull Lake ^{10}Be age (95-130 ka based on mixed cosmogenic ages, largely MIS 6), most of the glacial deposits of the Yellowstone Ice Cap in the current Paradise Valley, carved by the Yellowstone outlet glacier, are of Pinedale age (14-16.5ka based on ^{10}Be ages, MIS 2) (Pierce, 2004). Here, the ice extended further down valley during the Pinedale glaciation than during the Bull Lake glaciation due to uplift from the moving Yellowstone hotspot (Pierce, 2004). The ages of the Eightmile terminal moraines of the Yellowstone outlet glacier, 16.5-16.2 ka, are about 3000-4000 ^{10}Be years younger than moraines further south in the Rocky Mountains, possibly due

either to the time needed to build up the Yellowstone icecap or the climatic effect of the Laurentide ice sheet (Pierce, 2004). The moraines that underlie the Blacktail Deer Plateau up valley of the Blacktail pond have a cosmogenic age of 14.2 ± 1.2 ka, consistent with the Deckard Flats readjustment (Pierce et al., 2014), and 15.3 ± 1.4 ka down valley (Licciardi and Pierce, 2008), indicating that Blacktail Pond formed between 15,000 and 14,000 years ago (Krause and Whitlock, 2013) during the retreat of the Yellowstone outlet glacier. It would have formed in the wake of the rapid widespread deglaciation starting ~ 17 ka in the Rocky Mountains at the end of the last glacial in response to increasing summer insolation and greenhouse gas levels, and shifts in winter storm tracks (Clark and Bartlien, 1995; Licciardi et al., 2004). As there is no Younger Dryas signal in the sediment records of the area (12.9 to 11.5 cal ky BP; Alley et al., 2002) (Krause and Whitlock, 2013), ice recession continued through the summer insolation maximum at 9ka (Bartlien et al., 1998).

At the time of the LGM, the northern Rockies experienced fairly cold summer and dry winter conditions as a result of low greenhouse gas concentrations, low summer insolation, and the easterly anticyclonic flow off of the Laurentide ice sheet, as the jet stream was split and shifted south, weakening westerly air flow in both winter and summer and producing very cold temperatures over and adjacent to the ice sheet (Bartlein et al., 1998; Hostetler and Clark 1997). Though conditions were markedly different than those today, the pattern of maximum ELAs matches the pattern of modern ELAs, possibly indicating the same regional climatic differences exist today (Lock, 1990; Pierce, 2004). The northwest had dry summers, while and the southwest had wet

summers and mild winters (Mock and Bartlein, 1995). Plant species, primarily alpine tundra and meadow communities, with some Englemann spruce, supalpine fire, whitebark pine, lodgepole pine, and Douglas fir, existed in small population in species-specific settings near the ice margin and in non-glaciated valleys (Whitlock and Bartlein, 1993; Bartlein et al., 1998).

Holocene Paleoenvironmental History

During the late glacial, valley glaciers and large ice fields extended throughout the northern Rocky Mountains, notably including the Yellowstone Ice Cap. By about 17ka (Clark and Bartlein, 1995), rising summer insolation and greenhouse gases and shifts in winter storm tracks caused widespread deglaciation, made more pronounced after the summer insolation maximum at about 9ka (Clark and Bartlein 1995). The degree of warming was dramatic, similar to projections for the coming century (IPCC, 2007; USCCP, 2009). There were a series of step-like changes and climate reversals, including the Younger Dryas chronozone. There is little evidence of the Y.D. in Yellowstone, however, except for some possible glacial re-advances (Pierce, 2004).

Plant populations easily colonized new terrain following ice retreat (Whitlock and Bartlein, 1993). Similar patterns of colonization existed in all regions of Yellowstone. The sparse vegetation was followed by spruce parkland, then, as temperature rose, closed subalpine forest. Fire activity increased as bio mass increased and summers grew warmer (Bartlein et al., 1998). Increased summer insolation intensified circulation patterns, so seasonality was more intense as the ice sheet decreased in size. Summers became warmer and effectively drier, while lower winter insolation made winters colder and wetter due to

the northward shift of the jet stream (Bartlein et al, 1998), the decrease in winter insolation was less dramatic than the increase in the summer, indicating this decrease had less of an effect on climate (Figure 2.6) (Mock and Bartlein, 1995). The timing of maximum warmth and aridity during the Holocene varied within Yellowstone, especially after the summer insolation maximum at 9ka (Whitlock and Bartlein, 1993; Mock and Bartlein, 1995). The areas that are summer-wet today were wetter at 9ka and areas that are summer-dry today were also drier at 9ka (Mock and Bartlein, 1995).

Southern Yellowstone (summer-dry) experienced a peak in aridity and warmth between 9 and 5ka as a result of the expanded subtropical high, with wetter conditions in the late Holocene as summer insolation gradually dropped and the subtropical high weakened. In the early Holocene, lower elevation species like Douglas-fir and lodgepole pine forests developed until the middle to late Holocene, when summer drought was less severe as the subtropical high weakened, allowing for the growth of mesophytic conifers (Whitlock and Bartlein, 1993; Mock and Bartlein, 1995).

Between 9.5 and 7 ka, northern Yellowstone (summer-wet) experienced wetter than modern conditions, when monsoonal circulation was stronger, overwhelming the increased evapotranspiration. It experienced drying and warming after 7ka as the onshore flow from the Gulf of Mexico and Gulf of California weakened. (Whitlock and Bartlein, 1993, Mock and Bartlein, 1995). Lake levels lowered and grassland productivity increased due the increasingly arid conditions. Douglas-fir populations and fire activity increased (Whitlock and Bartlein, 1993). Studies of Blacktail Pond (Huerta et al., 2009; Krause and Whitlock, 2013), Crevice Lake (Whitlock et al., 2012), Dailey Lake (Krause

et al., 2015) and Slough Creek Pond (Whitlock and Bartlein, 1993; Millspaugh et al., 2000) add additional insight to local conditions.

Blacktail Pond is located in the summer-wet regime (Krause and Whitlock, 2013) on in an abandoned outwash channel (Pierce, 1979). When the lake formed, at about 14.6 ka in the wake of glacial retreat, climatic conditions were cool and it was populated by an alpine tundra community, and later by open spruce parkland. There was low fire activity due to cool conditions and little fuel. Rising insolation caused step-wise increases in temperatures, especially during the summer, and lower effective moisture (Figure B1) (Krause and Whitlock, 2013). The fact that it is a closed-basin lake means that $\delta^{18}\text{O}$ levels record summer temperature. Increasing $\delta^{18}\text{O}$ levels before 11.5ka support evidence for a transition to warmer summers. Increasing charcoal levels suggest fire activity increased due to increased fuel and increasingly warmer and drier summer conditions (Figure B1) (Krause and Whitlock, 2013). Additionally, immediately post-glacial, the lake showed little biological activity and a high influx of allochthonous sediment as evidenced by low Ca:Ti ratios and high magnetic susceptibility and inorganic silt and clay, representing little soil development in an unstable landscape yet to be stabilized by vegetation. Ca:Ti ratios increase and magnetic susceptibility decreases dramatically in response to further colonization by vegetation represented by the shift from tundra to parkland, and later a mixed conifer forest.

In the Early Holocene, 11.5-8.2ka, summer insolation reached its maximum directly increasing summer temperatures and indirectly strengthening the subtropical high and monsoonal systems (Bartlein et al., 1998). At this time, summers at Blacktail Pond

were hotter than present as evidenced by increasing $\delta^{18}\text{O}$. Nearby Crevice Lake, an open lake connected to the Yellowstone River, has $\delta^{18}\text{O}$ levels that represent winter precipitation. This record shows low $\delta^{18}\text{O}$ levels increasing supporting the idea that winter precipitation increased. The warm summers and wet winters led to a transition to a closed subalpine forest. Total pollen peaked at this time, evidencing an increase in total vegetation. This also indicates spring temperatures likely increased. Dense, dry fuels and high summer temperatures led to an increase in fire activity, allowing for the colonization of fire-adapted lodgepole pine, while fire-sensitive species like spruce, disappear from the record. Productivity was very high as Ca:Ti ratios peaked. It is worth noting there is no evidence of a Younger Dryas cold interval signal in $\delta^{18}\text{O}$ levels (Krause and Whitlock, 2013).

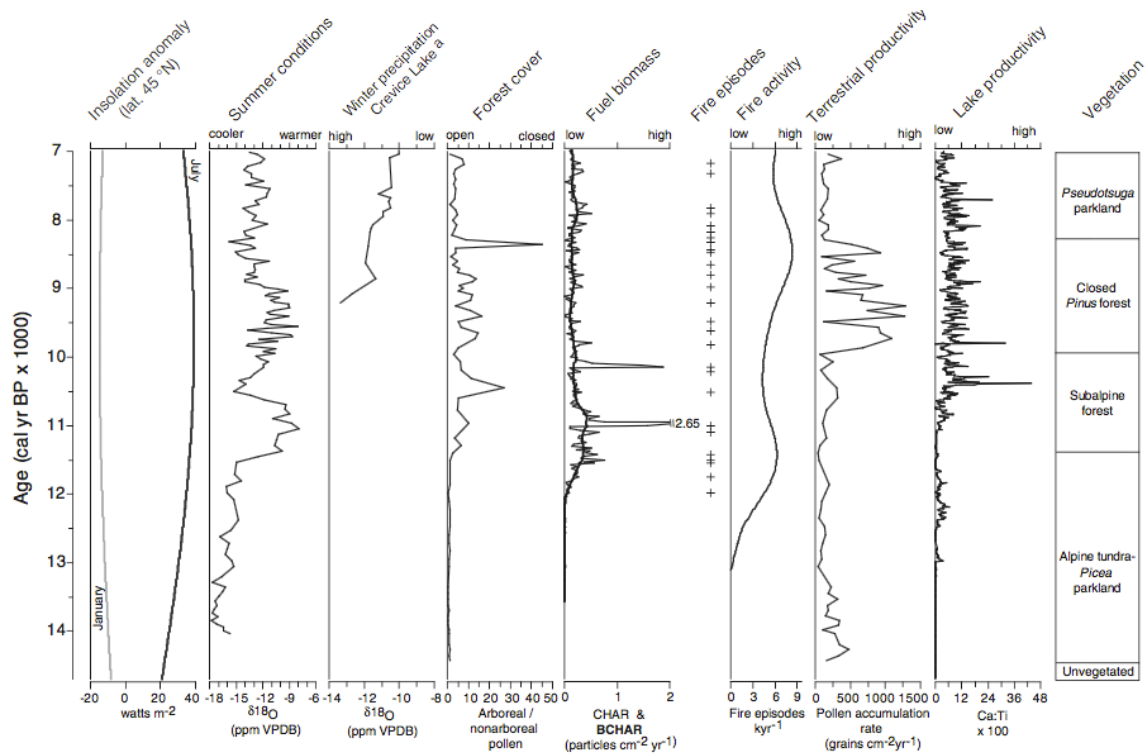


Figure B1 Paleoenvironmental data from Blacktail Pond from 14.6ka to 7ka. These are compared to January and July insolation. Figure from Krause and Whitlock, 2013. Insolation anomalies from Whitlock et al. 2012.

In the Middle Holocene, between 8.2 and 7ka, summer insolation decreased, resulting in a weakening monsoonal system. Winter insolation increased. Summers became cooler and drier and winters became milder with a lower snowpack. $\delta^{18}\text{O}$ levels in Blacktail Pond lowered, supporting cooler summers (Figure B1) (Krause and Whitlock, 2013). At Crevice Lake they increased, suggesting more winter precipitation was delivered as rain than snow (Whitlock, 2012). Total pollen levels decreased and the percentage of sage and Douglas fir pollen increased also suggesting drier summers and mild winters. Fire activity peaked at around 8ka in both northern and southern Yellowstone. This level in activity decreased in the south and central portions of

Yellowstone, but remained high in the northern areas. Northern Yellowstone continued to dry after 7ka (Figure B1) (Krause and Whitlock, 2013; Whitlock, 2012).

The characteristics of ^{10}Be suggest that those meteoric ^{10}Be concentrations in accumulated sediments should indicate changes in aspects of hydroclimate, specifically precipitation, and landscape stability. Meteoric ^{10}Be concentrations in the sediments of Blacktail Pond should reflect and add new insight to this rich record of environmental history in the Greater Yellowstone Ecosystem.

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APPENDIX C

10BE/9BE

As the versatility of meteoric ^{10}Be and its potential as a tracer in a wide variety of settings, as compared to in-situ produced ^{10}Be , has been investigated, meteoric ^{10}Be has not only been measured in past and present lake sediments (Belmaker et al., 2008; Horiuchi et al., 1999), but in ocean sediments (Bourles et al., 1989; Frank et al., 1997; Christl et al., 2003), river and marine terraces, (Pavich et al., 1986; Maejima et al., 2005), soil (Pavich et al., 1984; Graly et al., 2010), and river sediments (Brown et al. 1988, You et al., 1988, Willenbring and von Blanckenburg 2010, von Blanckenburg et al., 2012, Whittmann et al., 2012). Despite the advantages and versatility of this nuclide, there are challenges that comprise the ease of its use, and so it was not used in our work.

Though atmospheric production and delivery mechanisms and rates are becoming increasingly well understood (Field et al., 2006; Heikkila et al., 2007; Heikkila et al., 2008), the deposition rate must be known, especially when estimating erosion rates using meteoric ^{10}Be . While for large catchments, atmospheric mixing and deposition models may be used (Field et al., 2006; Heikkila et al., 2007; Willenbring and von Blanckenburg 2010), estimating deposition for smaller catchments may be more difficult (Whittmann et al., 2012).

Additionally, the behavior of meteoric ^{10}Be on the surface may be problematic. The retentivity of the nuclide on soil or sediment is highly dependent on grain size, in which concentrations of ^{10}Be increase with decreasing grain size, a characteristic attributed to increasing surface area with decreasing grain size (Willenbring and von Blanckenburg). In fluvial settings this characteristic adds complexity as particles are sorted by size. ^{10}Be has a high partition coefficient of around 10^5 mL/g at pH 6 and

above, depending on environment, indicating it is relatively immobile (Graley et al., 2010). This allows meteoric ^{10}Be to be retained on soil particles or incorporated as a cation into clays, organic matter, or oxyhydroxides under most natural chemical and mineralogical boundary conditions. This means that the pH of the soil should be greater than pH 6 (or at least between pH 3-11) so that particle reactive beryllium hydroxides and organic complexes are not lost to solution or competing with aluminum, which is mobilized at pH 6, and other metals for adsorption sites on clay particles. (Willenbring and von Blanckenburg, 2010). If the pH of the soil is below 6, ^{10}Be is lost to solution and carried away (You et al., 1989; Aldahan et al., 1999), inhibiting its use in studies measuring it in soil and sediment. Both the grain size and pH dependence of ^{10}Be mean that mineralogical composition governs the amount of ^{10}Be in an environment as different minerals have different grain sizes and influence on pH (Aldahan et al., 1999). Furthermore, the hydrology of a soil or landscape influence ^{10}Be retention. Dissolved ^{10}Be will only be lost to groundwater if the soils in a given environment are exposed to sufficient water for mobilization. Runoff and water transport at the surface will also influence the retention of ^{10}Be at the surface (Willenbring and von Blanckenburg, 2010).

A way to assess the retention of ^{10}Be in a given environment is to compare the concentration of ^{10}Be to the concentration of another element. Some early attempts to do so focused on using $^{10}\text{Be}/\text{Fe}$ ratios (Helz and Valette-Silver, 1992). However, the chemical properties of ^9Be , a stable isotope found in many crustal rocks in concentrations of around ~2 ppm, are much more similar to those of ^{10}Be . Hence, recent studies have utilized $^{10}\text{Be}/^9\text{Be}$ ratios to understand ^{10}Be behavior and to estimate erosion rates and

weathering intensities (von Blanckenburg et al., 2012; Bacon et al., 2012; Whittman et al., 2012). ^9Be is found primarily in crustal rocks and is released upon the weathering of the parent material in which it was bound. It either binds to particles or remains dissolved in solution (von Blanckenburg et al., 2011). After its release, it mixes with ^{10}Be in a ratio characteristic to a given environment dependent on denudation rates, deposition rates of ^{10}Be , and weathering intensity (von Blanckenburg et al., 2012).

In soils, ^9Be concentrations have been used to quantify the loss of ^{10}Be from a soil profile, and hence, both understand ^{10}Be behavior and estimate soil age (Bacon et al., 2012; Barg et al., 1997). Mass balance analysis of ^9Be and ^{10}Be in the weathering zone can yield denudation rates and weathering intensities as the ratio of ^{10}Be : ^9Be is also the ratio of the flux of ^{10}Be to the surface to the flux of ^9Be from the parent material (von Blanckenburg et al., 2012). The method can be applied to an entire river basin when the flux of dissolved ^9Be and ^{10}Be collected from rivers is compared to the ^9Be contribution to the watershed system from weathering and the ^{10}Be contribution from precipitation (von Blanckenburg et al., 2012). In rivers, the $^{10}\text{Be}/^9\text{Be}$ ratios have been used to understand the contribution of ^9Be to the ocean from rivers (Brown et al., 1992) and to quantify basin wide denudation rates (von Blanckenburg et al., 2011; Whittmann et al., 2012). Here, ^{10}Be can be normalized to ^9Be to account for the influence of grain size dependency. Similarly, meteoric ^{10}Be retention can be approximated using loss of ^9Be to determine soil residence time more accurately than measurements of ^{10}Be alone (Bacon et al., 2012) since it can be used to normalize differences in ^{10}Be concentrations with particle size (Willenbring and von Blanckenburg, 2010).

$^{10}\text{Be}/^9\text{Be}$ has been measured in ocean sediments (Bourles et al., 1989; Brown et al., 1992; Blanckenburg et al., 1996; Frank et al., 2008; Willenbring and von Blanckenburg, 2010b). Here, ^{10}Be is thought to be primarily of direct atmospheric origin, and that of continental origin is unlikely to show in long-term (2-12 Mya) variations (Willenbring and von Blanckenburg, 2010b). ^9Be in the ocean is derived from continental erosion delivered by fluvial sources either on suspended particles or dissolved in water (Brown et al., 1992; von Blanckenburg et al., 1996; Frank et al., 2009). Particles carrying ^9Be are buried in marine after some time, but some of the ^9Be may be recycled back in to the water. The modes of transport that influence ^9Be distribution in the ocean also influence the distribution of ^{10}Be , so the $^{10}\text{Be}/^9\text{Be}$ ratio will be characteristic of a given ocean basin (Igel et al., 1999). The same scavenging processes act on both isotopes, so it is thought that variations in $^{10}\text{Be}/^9\text{Be}$ ratios in ocean basins are caused only by the changing input of ^9Be via terrigenous delivery (Igel et al., 1999).

The studies of $^{10}\text{Be}/^9\text{Be}$ ratios in ocean sediments suggest that the variation in these ratios does not reflect changes in the delivery rate of ^{10}Be in the atmosphere on time scales on the order of a hundred thousand years due to mixing in large ocean basins (von Blanckenburg et al., 1999; von Blanckenburg et al., 1996). Combined with the idea that comparatively little ^{10}Be is delivered to oceans sediments from terrestrial sources, it seems $^{10}\text{Be}/^9\text{Be}$ ratios in a lake system would represent something other than a measure of changing erosional delivery of ^9Be to the lake. In a smaller-scale system much more impacted by the erosion of the surrounding landscape and by dust delivery, ^{10}Be is likely to be delivered to the lake via not just precipitation (and runoff), but by erosion and dust

delivery. This means that $^{10}\text{Be}/^9\text{Be}$ ratios will not reflect terrigenous input, especially on a timescale of less than 20ka. Because the same terrigenous delivery and scavenging processes act on the isotope, the $^{10}\text{Be}/^9\text{Be}$ ratios in the sediments of a small, closed basin lake will likely simply reflect the changing $^{10}\text{Be}/^9\text{Be}$ ratios of the sediment being delivered to the lake during different time periods. Studies of $^{10}\text{Be}/^9\text{Be}$ ratios in lake sediments have been oriented towards dating sediments in order to normalize ^{10}Be ages for grain size (Lebatard et al., 2010).

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APPENDIX D

CURRENT PALEOENVIRONMENTAL PROXIES

Several paleoenvironmental proxies are currently used in paleolimnology. A chronology of the sediments can be established with AMS radio carbon, and together with other paleoenvironmental proxies such as charcoal, fossil pollen, stable isotope data, and geochemical composition, we can infer such characteristics as fire, vegetation, and hydrologic history, as well as aspects of geologic setting and landscape stability. Meteoric ^{10}Be has the potential to be added to the suite of proxies to act as an indicator of changing hydroclimate and landscape stability.

Chronology

AMS ^{14}C ages may be determined from a combination of charcoal, plant microfossils, and pollen with a sediment record. In some records, such as the ones in Yellowstone, tephra and ash layers from volcanic eruptions many help constrain ages. The Blacktail Pond age model is based on seven AMS ^{14}C dates, three from Huerta et al., 2009 and four from Krause and Whitlock, 2013 as well as the Mazama ash layer at 6.7 ^{14}C BP (Zdanowicz et al., 1999) and the Glacier Peak ash layers at 11.6 ^{14}C BP (Kuehn et al., 2009) identified in the core. This dating allowed multi-decadal resolution.

Stable Isotopes

$\delta^{18}\text{O}$ values in authigenic carbonates in a sediment core provide insight into the climate at the time the carbonate was deposited. Authigenic carbonate is precipitated via photosynthetic utilization of CO_2 by planktonic organisms and algae. Their consumption of CO_2 results in higher pH and carbonate levels until carbonate supersaturates the water and binds to available Ca^{2+} . Levels of calcite are often used as a proxy for productivity

and temperature since photosynthesis occurs with more sunlight and higher temperatures (Meyers and Ishiwatari, 1993).

$\delta^{18}\text{O}$ represents aspects of temperature, hydrology, and the timing of high productivity. Generally, increased $\delta^{18}\text{O}$ levels mean increased rainfall, decreased snowfall, more rain to snow or high evaporation, but the dominance of each factor is dependent on the setting or hydrology of the lake (Leng and Marshall, 2004). In small lakes and open lakes with inflow and outflow or a significant groundwater connection, small variations in $\delta^{18}\text{O}$ represent variations in hydrology, the isotopic composition of precipitation, or the seasonality and the type of precipitation delivered. Here, low values often reflect more snow and/or less rain, though they may also be influenced by evaporation (Leng and Marshall, 2004; Anderson, 2011). In larger lakes or closed basin lakes, large $\delta^{18}\text{O}$ variations usually represent changes in temperature, and hence evaporation, reflecting a long-term precipitation to evaporation ratio. In this kind of system, higher values represent more evaporation and higher summer temperatures. However, they may also represent increased rainfall and less snow (Leng and Marshall, 2004; Anderson, 2011). Additionally, the influence of groundwater or a change in precipitation source cannot be ruled out. While $\delta^{18}\text{O}$ provides a useful proxy for past hydroclimate, its concentrations are also influenced by other factors, meaning that the $\delta^{18}\text{O}$ signal is the product of a number of combined complex processes. In addition to the complexities associated with hydrology, climate, and geography, there are chemical and biogenic effects that cause fractionation differences that might not be predicted by thermodynamics. These depend on which carbonate producing species are present in a

lake, the conditions under which these species thrive, and the biogenic pathways they utilize to produce carbonate. Understanding authogenically derived $\delta^{18}\text{O}$ may also prove more complicated if the lake sits on a substrate with a high carbonate content. If $\delta^{18}\text{O}$ is used in conjunction with diatom or pollen records it provides a robust integrated climate signal (Leng and Marshall, 2004).

Pollen

Fossilized pollen is used to reconstruct vegetation history. Comparison of pollen types found in sediments to modern pollen helps to identify the vegetation species alive over the span of the record (Whitlock, 1993; Krause and Whitlock, 2013). Additionally, ratios of pollen species allow for the reconstruction of vegetative communities and pollen accumulation rates and abundance help reconstruct vegetation density and abundance (Krause and Whitlock, 2013). In turn, this combined information aide in climatic reconstruction as different species and assemblages of species require different climatic conditions (Krause and Whitlock, 2013).

Charcoal

Particles of charcoal are used to reconstruct different aspects of fire activity (Whitlock and Millspaugh, 1995). Charcoal analysis involves the quantification of accumulated charcoal particles in sediments, correlating them to a fire event. The charcoal can be radiocarbon dated, and sedimentation rates can be calculated (Marlon et al. 2005). Charcoal production, transport, and deposition depend on the size, severity, and intensity of a fire. Though it can be difficult to infer location of a fire, the size of charcoal

particles can be a clue, as smaller particles tend to travel further. (Whitlock and Larsen, 2001). Because charcoal particles can be carried long distances and may be transported years after a fire, sediment records tend to reflect “fire episodes” or events in the general area, rather than individual fires (Whitlock and Barlein, 2004). Thus, changes in Holocene fire regimes can be determined. Long-term charcoal accumulation rates represent general long-term biomass. Peaks in rate represent the fire return interval (Krause and Whitlock, 2013). On long timescales, climate is the most significant control of fire regime. Changes in insolation, atmospheric composition, and ocean-atmosphere exchange, influence both global and regional climate patterns, which in turn, affect, regional moisture, temperature, and dryness, factors that directly influence fire frequency (Whitlock and Bartlein, 2004). “Fire regime” is defined on the basis of a combination of fuel type, temporal characteristics (e.g. frequency), spatial characteristics (e.g. size), and consequences (i.e. the impact of intensity on vegetation and soils) (Bowman et al. 2009). When sedimentary charcoal records are coupled with climate models and proxy that reveal boundary conditions, researchers gain insight on climatic, as well as anthropogenic, forcing on fire frequency.

Geochemistry

The geochemical composition of the core reflects a number of parameters. It can help uncover the chemical processes taking place in the lake and the type of material delivered to the lake throughout the record as well as the proportion of allochthonous to autochthonous sediment.

Magnetic Susceptibility

Magnetic iron species are an indicator of detrital origin (Gedye et al 2000; Demory et al., 2005). Studies of sediment records measure magnetic susceptibility, which is a dimensionless proportionality constant that indicates the degree to magnetization of a mineral in response to an applied magnetic field (Anderson and Hallet, 1996). It is represented by χ and defined in S.I. units (Anderson and Hallet, 1996). While iron is fairly mobile and involved in the redox history of sediments, most magnetic assemblages in lakes sediments are derived from terrigenous input and often show strong similarity to titanium curves obtained from the same core (Demory et al., 2005).

Ca:Ti

In paleoclimate studies, carbonate production is often used as an indicator of overall lake productivity since the photosynthesis of macroscopic plants or algae consumes CO₂, which is thought to increase carbonate ion activity in the water. Carbonate precipitation is triggered by algal photosynthesis in the summer. The carbonate binds to available Ca²⁺ and precipitates as calcium carbonate (Leng and Marshall, 2004). Rates of photosynthesis are higher with warmer and sunnier conditions, so higher rates of calcite precipitation is thought to be associated with such conditions during summer months in a given setting (Meyers and Ishiwarari, 1993). Since calcium enters the lake system via both biogenic and weathering and erosion processes, quantifying the ratio of Ca to Ti, provides an estimate of how much calcium was produced in the lake, versus calcium that was introduced to the lake from terrestrial sources because titanium is allocthonous and immobile. Titanium is a detrital sediment indicator because it is only produced

allogenicly through the erosion of Ti bearing rocks (Cohen, 2003). Ti is immobile and insensitive to dissolution (Demory, et al, 2005), meaning that it can consistently act as a measure of terrigenous input to the lake. Since calcium enters the lake system via both biogenic and weathering and erosion processes, quantifying the ratio of Ca to Ti, provides an estimate of how much calcium was produced in the lake, versus calcium that was introduced to the lake from terrestrial sources.

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APPENDIX E

BLACKTAIL POND DATA USED IN MODEL

Equations 1-4 in Chapter 3, Appendix 3A utilize variables from Krause and Whitlock (2013) to predict ^{10}Be concentrations with depth in the Blacktail Pond core. Depths are reported every 0.25cm with their corresponding age obtained from the Krause and Whitlock (2013) age-depth model. These comprise the first two columns of the below table. Krause and Whitlock (2013) also obtained XRF data from the Blacktail Pond core, and resulting titanium counts are shown in the last column of this table. The third column of this table shows constant modern precipitation in Yellowstone (38.00 cm/yr). The fourth and fifth column are comprised of the values used to predict ^{10}Be flux in the model from Graly et al. (2011), showing a peak in precipitation at 9ka that is 10% higher than modern precipitation or 50% higher than modern. These precipitation values are also plotted with depth in Figure 3.15.

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
525.601	6984.558	38.000	38.000	38.000	642.000
525.851	6993.347	38.000	38.002	38.017	720.000
526.101	7002.143	38.000	38.004	38.033	778.000
526.351	7010.947	38.000	38.006	38.050	840.000
526.601	7019.759	38.000	38.007	38.067	741.000
526.851	7028.578	38.000	38.009	38.083	732.000
527.101	7037.405	38.000	38.011	38.100	635.000
527.351	7046.240	38.000	38.013	38.117	737.000
527.601	7055.083	38.000	38.015	38.133	997.000
527.851	7063.934	38.000	38.017	38.150	932.000
528.101	7072.792	38.000	38.019	38.167	845.000
528.351	7081.659	38.000	38.020	38.183	542.000
528.601	7090.533	38.000	38.022	38.200	397.000
528.851	7099.414	38.000	38.024	38.217	379.000
529.101	7108.304	38.000	38.026	38.233	354.000
529.351	7117.201	38.000	38.028	38.250	274.000
529.601	7126.106	38.000	38.030	38.267	405.000
529.851	7135.019	38.000	38.031	38.283	399.000
530.101	7143.940	38.000	38.033	38.300	430.000
530.351	7152.869	38.000	38.035	38.317	432.000
530.601	7161.805	38.000	38.037	38.333	344.000
530.851	7170.750	38.000	38.039	38.350	238.000
531.101	7179.702	38.000	38.041	38.367	245.000
531.351	7188.662	38.000	38.043	38.383	251.000
531.601	7197.630	38.000	38.044	38.400	355.000
531.851	7206.605	38.000	38.046	38.417	565.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
532.351	7224.580	38.000	38.050	38.450	417.000
532.601	7233.579	38.000	38.052	38.467	451.000
532.851	7242.587	38.000	38.054	38.483	255.000
533.101	7251.601	38.000	38.056	38.500	454.000
533.351	7260.624	38.000	38.057	38.517	490.000
533.601	7269.655	38.000	38.059	38.533	412.000
533.851	7278.694	38.000	38.061	38.550	756.000
534.101	7287.740	38.000	38.063	38.567	1378.000
534.351	7296.794	38.000	38.065	38.583	1946.000
534.601	7305.856	38.000	38.067	38.600	1835.000
534.851	7314.927	38.000	38.069	38.617	1537.000
535.101	7324.004	38.000	38.070	38.633	1345.000
535.351	7333.090	38.000	38.072	38.650	876.000
535.601	7342.184	38.000	38.074	38.667	507.000
535.851	7351.286	38.000	38.076	38.683	533.000
536.101	7360.395	38.000	38.078	38.700	526.000
536.351	7369.513	38.000	38.080	38.717	520.000
536.601	7378.638	38.000	38.081	38.733	711.000
536.851	7387.772	38.000	38.083	38.750	583.000
537.101	7396.913	38.000	38.085	38.767	501.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
537.601	7415.219	38.000	38.089	38.800	1008.000
537.851	7424.384	38.000	38.091	38.817	1168.000
538.101	7433.557	38.000	38.093	38.833	984.000
538.351	7442.738	38.000	38.094	38.850	883.000
538.601	7451.927	38.000	38.096	38.867	660.000
538.851	7461.123	38.000	38.098	38.883	492.000
539.101	7470.328	38.000	38.100	38.900	407.000
539.351	7479.541	38.000	38.106	38.921	402.000
539.601	7488.762	38.000	38.113	38.942	516.000
539.851	7497.990	38.000	38.119	38.963	385.000
540.101	7507.227	38.000	38.125	38.984	426.000
540.351	7516.471	38.000	38.132	39.005	509.000
540.601	7525.724	38.000	38.138	39.025	731.000
540.851	7534.984	38.000	38.145	39.046	882.000
541.101	7544.253	38.000	38.151	39.067	965.000
541.351	7553.529	38.000	38.157	39.088	759.000
541.601	7562.814	38.000	38.164	39.109	932.000
541.851	7572.106	38.000	38.170	39.130	835.000
542.101	7581.407	38.000	38.176	39.151	1386.000
542.351	7590.715	38.000	38.183	39.172	2150.000
542.601	7600.031	38.000	38.189	39.193	2750.000
542.851	7609.356	38.000	38.195	39.214	3344.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
543.351	7628.029	38.000	38.208	39.255	1084.000
543.601	7637.377	38.000	38.215	39.276	734.000
543.851	7646.734	38.000	38.221	39.297	467.000
544.101	7656.098	38.000	38.227	39.318	495.000
544.351	7665.469	38.000	38.234	39.339	554.000
544.601	7674.848	38.000	38.240	39.360	657.000
544.851	7684.232	38.000	38.246	39.381	520.000
545.101	7693.622	38.000	38.253	39.402	298.000
545.351	7703.017	38.000	38.259	39.423	197.000
545.601	7712.416	38.000	38.265	39.444	335.000
545.851	7721.819	38.000	38.272	39.465	232.000
546.101	7731.225	38.000	38.278	39.485	256.000
546.351	7740.634	38.000	38.285	39.506	280.000
546.601	7750.046	38.000	38.291	39.527	196.000
546.851	7759.459	38.000	38.297	39.548	522.000
547.101	7768.873	38.000	38.304	39.569	618.000
547.351	7778.287	38.000	38.310	39.590	662.000
547.601	7787.701	38.000	38.316	39.611	513.000
547.851	7797.115	38.000	38.323	39.632	401.000
548.101	7806.528	38.000	38.329	39.653	454.000
548.351	7815.938	38.000	38.335	39.674	387.000
548.601	7825.347	38.000	38.342	39.695	228.000
548.851	7834.752	38.000	38.348	39.715	208.000
549.101	7844.154	38.000	38.355	39.736	459.000
549.351	7853.552	38.000	38.361	39.757	427.000
549.601	7862.946	38.000	38.367	39.778	514.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
550.101	7881.717	38.000	38.380	39.820	315.000
550.351	7891.093	38.000	38.386	39.841	206.000
550.601	7900.463	38.000	38.393	39.862	474.000
550.851	7909.825	38.000	38.399	39.883	258.000
551.101	7919.179	38.000	38.405	39.904	471.000
551.351	7928.525	38.000	38.412	39.925	404.000
551.601	7937.862	38.000	38.418	39.945	379.000
551.851	7947.188	38.000	38.425	39.966	508.000
552.101	7956.505	38.000	38.431	39.987	666.000
552.351	7965.811	38.000	38.437	40.008	573.000
552.601	7975.106	38.000	38.444	40.029	435.000
552.851	7984.389	38.000	38.450	40.050	353.000
553.101	7993.659	38.000	38.450	40.070	370.000
553.351	8002.917	38.000	38.450	40.090	110.000
553.601	8012.160	38.000	38.451	40.111	241.000
553.851	8021.390	38.000	38.451	40.131	307.000
554.101	8030.605	38.000	38.451	40.151	190.000
554.351	8039.805	38.000	38.451	40.171	277.000
554.601	8048.988	38.000	38.451	40.191	393.000
554.851	8058.156	38.000	38.451	40.211	393.000
555.101	8067.306	38.000	38.452	40.232	318.000
555.351	8076.439	38.000	38.452	40.252	487.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
555.851	8094.651	38.000	38.452	40.292	543.000
556.101	8103.728	38.000	38.452	40.312	349.000
556.351	8112.785	38.000	38.453	40.333	243.000
556.601	8121.822	38.000	38.453	40.353	540.000
556.851	8130.838	38.000	38.453	40.373	290.000
557.101	8139.833	38.000	38.453	40.393	375.000
557.351	8148.805	38.000	38.453	40.413	498.000
557.601	8157.756	38.000	38.453	40.433	228.000
557.851	8166.682	38.000	38.454	40.454	184.000
558.101	8175.586	38.000	38.454	40.474	246.000
558.351	8184.465	38.000	38.454	40.494	297.000
558.601	8193.319	38.000	38.454	40.514	247.000
558.851	8202.148	38.000	38.454	40.534	239.000
559.101	8210.951	38.000	38.455	40.555	254.000
559.351	8219.727	38.000	38.455	40.575	396.000
559.601	8228.476	38.000	38.455	40.595	408.000
559.851	8237.198	38.000	38.455	40.615	288.000
560.101	8245.891	38.000	38.455	40.635	309.000
560.351	8254.556	38.000	38.455	40.655	309.000
560.601	8263.191	38.000	38.456	40.676	198.000
560.851	8271.797	38.000	38.456	40.696	290.000
561.101	8280.372	38.000	38.456	40.716	187.000
561.351	8288.916	38.000	38.456	40.736	141.000
561.601	8297.428	38.000	38.456	40.756	328.000
561.851	8305.908	38.000	38.457	40.777	175.000
562.101	8314.356	38.000	38.457	40.797	153.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
562.601	8331.151	38.000	38.457	40.837	438.000
562.851	8339.497	38.000	38.457	40.857	455.000
563.101	8347.808	38.000	38.457	40.877	389.000
563.351	8356.084	38.000	38.458	40.898	268.000
563.601	8364.323	38.000	38.458	40.918	219.000
563.851	8372.526	38.000	38.458	40.938	258.000
564.101	8380.692	38.000	38.458	40.958	253.000
564.351	8388.820	38.000	38.458	40.978	208.000
564.601	8396.910	38.000	38.459	40.999	333.000
564.851	8404.961	38.000	38.459	41.019	315.000
565.101	8412.974	38.000	38.459	41.039	244.000
565.351	8420.949	38.000	38.459	41.059	335.000
565.601	8428.886	38.000	38.459	41.079	294.000
565.851	8436.785	38.000	38.459	41.099	318.000
566.101	8444.647	38.000	38.460	41.120	384.000
566.351	8452.471	38.000	38.460	41.140	304.000
566.601	8460.259	38.000	38.460	41.160	130.000
566.851	8468.010	38.000	38.470	41.184	190.000
567.101	8475.725	38.000	38.480	41.209	280.000
567.351	8483.403	38.000	38.489	41.233	235.000
567.601	8491.045	38.000	38.499	41.257	205.000
567.851	8498.652	38.000	38.509	41.282	228.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
568.351	8513.758	38.000	38.529	41.331	232.000
568.601	8521.259	38.000	38.539	41.355	188.000
568.851	8528.725	38.000	38.548	41.379	181.000
569.101	8536.156	38.000	38.558	41.404	208.000
569.351	8543.552	38.000	38.568	41.428	295.000
569.601	8550.915	38.000	38.578	41.452	458.000
569.851	8558.243	38.000	38.588	41.477	341.000
570.101	8565.538	38.000	38.597	41.501	298.000
570.351	8572.800	38.000	38.607	41.525	357.000
570.601	8580.028	38.000	38.617	41.550	287.000
570.851	8587.223	38.000	38.627	41.574	354.000
571.101	8594.386	38.000	38.637	41.599	347.000
571.351	8601.515	38.000	38.647	41.623	359.000
571.601	8608.613	38.000	38.656	41.647	354.000
571.851	8615.678	38.000	38.666	41.672	356.000
572.101	8622.712	38.000	38.676	41.696	337.000
572.351	8629.714	38.000	38.686	41.720	326.000
572.601	8636.685	38.000	38.696	41.745	290.000
572.851	8643.624	38.000	38.705	41.769	281.000
573.101	8650.533	38.000	38.715	41.793	350.000
573.351	8657.411	38.000	38.725	41.818	326.000
573.601	8664.258	38.000	38.735	41.842	388.000
573.851	8671.076	38.000	38.745	41.867	327.000
574.101	8677.863	38.000	38.755	41.891	322.000
574.351	8684.620	38.000	38.764	41.915	301.000
574.601	8691.348	38.000	38.774	41.940	216.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
575.101	8704.716	38.000	38.794	41.988	149.000
575.351	8711.357	38.000	38.804	42.013	247.000
575.601	8717.969	38.000	38.813	42.037	314.000
575.851	8724.553	38.000	38.823	42.061	203.000
576.101	8731.109	38.000	38.833	42.086	166.000
576.351	8737.636	38.000	38.843	42.110	219.000
576.601	8744.137	38.000	38.853	42.135	286.000
576.851	8750.609	38.000	38.863	42.159	304.000
577.101	8757.055	38.000	38.872	42.183	238.000
577.351	8763.473	38.000	38.882	42.208	241.000
577.601	8769.865	38.000	38.892	42.232	272.000
577.851	8776.230	38.000	38.902	42.256	162.000
578.101	8782.569	38.000	38.912	42.281	160.000
578.351	8788.882	38.000	38.921	42.305	163.000
578.601	8795.169	38.000	38.931	42.329	276.000
578.851	8801.430	38.000	38.941	42.354	358.000
579.101	8807.666	38.000	38.951	42.378	332.000
579.351	8813.877	38.000	38.961	42.403	355.000
579.601	8820.063	38.000	38.971	42.427	360.000
579.851	8826.224	38.000	38.980	42.451	486.000
580.101	8832.361	38.000	38.990	42.476	362.000
580.351	8838.474	38.000	39.000	42.500	401.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
580.851	8850.628	38.000	39.029	42.651	474.000
581.101	8856.669	38.000	39.043	42.697	498.000
581.351	8862.688	38.000	39.057	42.743	370.000
581.601	8868.683	38.000	39.072	42.790	378.000
581.851	8874.655	38.000	39.086	42.836	235.000
582.101	8880.605	38.000	39.100	42.882	227.000
582.351	8886.532	38.000	39.115	42.929	202.000
582.601	8892.438	38.000	39.129	42.975	264.000
582.851	8898.321	38.000	39.144	43.021	234.000
583.101	8904.183	38.000	39.158	43.068	305.000
583.351	8910.023	38.000	39.172	43.114	304.000
583.601	8915.843	38.000	39.187	43.160	224.000
583.851	8921.641	38.000	39.201	43.207	163.000
584.101	8927.418	38.000	39.215	43.253	201.000
584.351	8933.175	38.000	39.230	43.299	230.000
584.601	8938.912	38.000	39.244	43.345	260.000
584.851	8944.629	38.000	39.258	43.392	220.000
585.101	8950.326	38.000	39.273	43.438	172.000
585.351	8956.003	38.000	39.287	43.484	390.000
585.601	8961.661	38.000	39.301	43.531	330.000
585.851	8967.300	38.000	39.316	43.577	196.000
586.101	8972.920	38.000	39.330	43.623	319.000
586.351	8978.522	38.000	39.344	43.670	257.000
586.601	8984.105	38.000	39.359	43.716	256.000
586.851	8989.670	38.000	39.373	43.762	372.000
587.101	8995.216	38.000	39.387	43.809	344.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
587.601	9006.257	38.000	39.416	43.901	158.000
587.851	9011.752	38.000	39.431	43.948	258.000
588.101	9017.229	38.000	39.445	43.994	257.000
588.351	9022.690	38.000	39.459	44.040	173.000
588.601	9028.134	38.000	39.474	44.087	291.000
588.851	9033.562	38.000	39.488	44.133	257.000
589.101	9038.973	38.000	39.502	44.179	251.000
589.351	9044.369	38.000	39.517	44.226	350.000
589.601	9049.749	38.000	39.531	44.272	495.000
589.851	9055.114	38.000	39.545	44.318	370.000
590.101	9060.464	38.000	39.560	44.365	410.000
590.351	9065.798	38.000	39.574	44.411	220.000
590.601	9071.118	38.000	39.588	44.457	288.000
590.851	9076.424	38.000	39.603	44.504	209.000
591.101	9081.715	38.000	39.617	44.550	216.000
591.351	9086.992	38.000	39.631	44.596	397.000
591.601	9092.256	38.000	39.646	44.642	389.000
591.851	9097.505	38.000	39.660	44.689	353.000
592.101	9102.742	38.000	39.675	44.735	293.000
592.351	9107.966	38.000	39.689	44.781	245.000
592.601	9113.176	38.000	39.703	44.828	349.000
592.851	9118.374	38.000	39.718	44.874	323.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
593.351	9128.733	38.000	39.746	44.967	227.000
593.601	9133.895	38.000	39.761	45.013	287.000
593.851	9139.045	38.000	39.775	45.000	242.000
594.101	9144.183	38.000	39.786	45.045	182.000
594.351	9149.311	38.000	39.800	45.091	155.000
594.601	9154.427	38.000	39.814	45.136	101.000
594.851	9159.532	38.000	39.828	45.182	238.000
595.101	9164.627	38.000	39.842	45.227	179.000
595.351	9169.712	38.000	39.856	45.273	222.000
595.601	9174.786	38.000	39.871	45.318	134.000
595.851	9179.851	38.000	39.885	45.364	200.000
596.101	9184.906	38.000	39.899	45.409	203.000
596.351	9189.952	38.000	39.913	45.455	196.000
596.601	9194.988	38.000	39.927	45.500	307.000
596.851	9200.016	38.000	39.941	45.545	322.000
597.101	9205.035	38.000	39.955	45.591	376.000
597.351	9210.046	38.000	39.969	45.636	363.000
597.601	9215.048	38.000	39.983	45.682	198.000
597.851	9220.042	38.000	39.997	45.727	332.000
598.101	9225.029	38.000	40.011	45.773	370.000
598.351	9230.008	38.000	40.026	45.818	449.000
598.601	9234.979	38.000	40.040	45.864	703.000
598.851	9239.944	38.000	40.054	45.909	737.000
599.101	9244.902	38.000	40.068	45.955	571.000
599.351	9249.853	38.000	40.082	46.000	450.000
599.601	9254.798	38.000	40.096	46.045	387.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
600.101	9264.670	38.000	40.124	46.136	252.000
600.351	9269.597	38.000	40.138	46.182	232.000
600.601	9274.519	38.000	40.152	46.227	252.000
600.851	9279.435	38.000	40.166	46.273	266.000
601.101	9284.347	38.000	40.181	46.318	410.000
601.351	9289.254	38.000	40.195	46.364	375.000
601.601	9294.156	38.000	40.209	46.409	287.000
601.851	9299.054	38.000	40.223	46.455	347.000
602.101	9303.948	38.000	40.237	46.500	274.000
602.351	9308.838	38.000	40.251	46.545	219.000
602.601	9313.725	38.000	40.265	46.591	354.000
602.851	9318.608	38.000	40.279	46.636	211.000
603.101	9323.488	38.000	40.293	46.682	215.000
603.351	9328.366	38.000	40.307	46.727	450.000
603.601	9333.240	38.000	40.321	46.773	271.000
603.851	9338.112	38.000	40.336	46.818	362.000
604.101	9342.982	38.000	40.350	46.864	251.000
604.351	9347.850	38.000	40.364	46.909	155.000
604.601	9352.717	38.000	40.378	46.955	221.000
604.851	9357.581	38.000	40.392	47.000	199.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
605.351	9367.307	38.000	40.420	47.091	294.000
605.601	9372.169	38.000	40.434	47.136	317.000
605.851	9377.030	38.000	40.448	47.182	243.000
606.101	9381.891	38.000	40.462	47.227	180.000
606.351	9386.751	38.000	40.476	47.273	173.000
606.601	9391.612	38.000	40.491	47.318	176.000
606.851	9396.473	38.000	40.505	47.364	204.000
607.101	9401.335	38.000	40.519	47.409	292.000
607.351	9406.197	38.000	40.533	47.455	406.000
607.601	9411.061	38.000	40.547	47.500	392.000
607.851	9415.926	38.000	40.561	47.545	521.000
608.101	9420.792	38.000	40.575	47.591	516.000
608.351	9425.660	38.000	40.589	47.636	361.000
608.601	9430.530	38.000	40.603	47.682	199.000
608.851	9435.402	38.000	40.617	47.727	260.000
609.101	9440.277	38.000	40.631	47.773	197.000
609.351	9445.155	38.000	40.646	47.818	195.000
609.601	9450.035	38.000	40.660	47.864	153.000
609.851	9454.918	38.000	40.674	47.909	157.000
610.101	9459.805	38.000	40.688	47.955	160.000
610.351	9464.695	38.000	40.702	48.000	219.000
610.601	9469.590	38.000	40.716	48.045	318.000
610.851	9474.488	38.000	40.730	48.091	210.000
611.101	9479.390	38.000	40.744	48.136	288.000
611.351	9484.297	38.000	40.758	48.182	445.000
611.601	9489.209	38.000	40.772	48.227	338.000
611.851	9494.126	38.000	40.786	48.273	325.000
612.101	9499.048	38.000	40.801	48.318	298.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
612.601	9508.909	38.000	40.829	48.409	137.000
612.851	9513.848	38.000	40.843	48.455	200.000
613.101	9518.794	38.000	40.857	48.500	236.000
613.351	9523.746	38.000	40.871	48.545	249.000
613.601	9528.705	38.000	40.885	48.591	264.000
613.851	9533.672	38.000	40.899	48.636	340.000
614.101	9538.645	38.000	40.913	48.682	444.000
614.351	9543.627	38.000	40.927	48.727	426.000
614.601	9548.617	38.000	40.941	48.773	537.000
614.851	9553.615	38.000	40.956	48.818	494.000
615.101	9558.622	38.000	40.970	48.864	382.000
615.351	9563.637	38.000	40.984	48.909	303.000
615.601	9568.662	38.000	40.998	48.955	231.000
615.851	9573.696	38.000	41.012	49.000	137.000
616.101	9578.741	38.000	41.026	49.045	227.000
616.351	9583.795	38.000	41.040	49.091	203.000
616.601	9588.860	38.000	41.054	49.136	236.000
616.851	9593.935	38.000	41.068	49.182	242.000
617.101	9599.021	38.000	41.082	49.227	278.000
617.351	9604.119	38.000	41.096	49.273	274.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
617.851	9614.348	38.000	41.125	49.364	185.000
618.101	9619.481	38.000	41.139	49.409	151.000
618.351	9624.627	38.000	41.153	49.455	230.000
618.601	9629.785	38.000	41.167	49.500	313.000
618.851	9634.956	38.000	41.181	49.545	424.000
619.101	9640.140	38.000	41.195	49.591	314.000
619.351	9645.338	38.000	41.209	49.636	399.000
619.601	9650.549	38.000	41.223	49.682	342.000
619.851	9655.775	38.000	41.237	49.727	381.000
620.101	9661.015	38.000	41.251	49.773	421.000
620.351	9666.270	38.000	41.266	49.818	417.000
620.601	9671.540	38.000	41.280	49.864	418.000
620.851	9676.825	38.000	41.294	49.909	186.000
621.101	9682.126	38.000	41.308	49.955	150.000
621.351	9687.442	38.000	41.325	50.000	176.000
621.601	9692.775	38.000	41.335	50.142	257.000
621.851	9698.124	38.000	41.344	50.285	223.000
622.101	9703.490	38.000	41.354	50.427	216.000
622.351	9708.873	38.000	41.364	50.569	303.000
622.601	9714.273	38.000	41.373	50.711	268.000
622.851	9719.691	38.000	41.383	50.854	227.000
623.101	9725.126	38.000	41.393	50.996	263.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
623.601	9736.052	38.000	41.412	51.280	318.000
623.851	9741.543	38.000	41.422	51.423	250.000
624.101	9747.053	38.000	41.431	51.565	280.000
624.351	9752.582	38.000	41.441	51.707	200.000
624.601	9758.131	38.000	41.451	51.850	242.000
624.851	9763.700	38.000	41.460	51.992	267.000
625.101	9769.289	38.000	41.470	52.134	18.000
625.351	9774.898	38.000	41.479	52.276	0.000
625.601	9780.528	38.000	41.489	52.419	28.000
625.401	9776.023	38.000	41.481	52.305	260.000
625.651	9781.657	38.000	41.491	52.447	308.000
625.901	9787.312	38.000	41.501	52.589	307.000
626.151	9792.989	38.000	41.510	52.732	327.000
626.401	9798.688	38.000	41.520	52.874	359.000
626.651	9804.408	38.000	41.530	53.016	302.000
626.901	9810.151	38.000	41.539	53.159	409.000
627.151	9815.916	38.000	41.549	53.301	579.000
627.401	9821.704	38.000	41.559	53.443	598.000
627.651	9827.516	38.000	41.568	53.585	462.000
627.901	9833.350	38.000	41.578	53.728	491.000
628.151	9839.209	38.000	41.588	53.870	612.000
628.401	9845.091	38.000	41.597	54.012	537.000
628.651	9850.997	38.000	41.607	54.154	494.000
628.901	9856.929	38.000	41.617	54.297	358.000
629.151	9862.885	38.000	41.626	54.439	396.000
629.401	9868.866	38.000	41.636	54.581	335.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
629.901	9880.905	38.000	41.655	54.866	304.000
630.151	9886.963	38.000	41.665	55.008	210.000
630.401	9893.048	38.000	41.674	55.150	144.000
630.651	9899.159	38.000	41.684	55.293	293.000
630.901	9905.297	38.000	41.694	55.435	313.000
631.151	9911.462	38.000	41.703	55.577	269.000
631.401	9917.655	38.000	41.713	55.720	334.000
631.651	9923.875	38.000	41.723	55.862	402.000
631.901	9930.123	38.000	41.732	56.004	653.000
632.151	9936.400	38.000	41.742	56.146	1576.000
632.401	9942.705	38.000	41.752	56.289	926.000
632.651	9949.040	38.000	41.761	56.431	589.000
632.901	9955.403	38.000	41.771	56.573	394.000
633.151	9961.796	38.000	41.781	56.715	333.000
633.401	9968.218	38.000	41.790	56.858	329.000
633.651	9974.671	38.000	41.800	57.000	247.000
633.901	9981.154	38.000	41.798	56.988	206.000
634.151	9987.668	38.000	41.797	56.976	320.000
634.401	9994.212	38.000	41.795	56.965	266.000
634.651	10000.788	38.000	41.793	56.953	322.000
634.901	10007.395	38.000	41.791	56.941	208.000
635.151	10014.034	38.000	41.790	56.929	186.000
635.401	10020.705	38.000	41.788	56.918	201.000
635.651	10027.408	38.000	41.786	56.906	305.000
635.901	10034.144	38.000	41.784	56.894	259.000
636.151	10040.913	38.000	41.783	56.882	228.000
636.401	10047.715	38.000	41.781	56.871	146.000
636.651	10054.551	38.000	41.779	56.859	98.000
636.901	10061.420	38.000	41.777	56.847	147.000
637.151	10068.324	38.000	41.776	56.835	151.000
637.401	10075.261	38.000	41.774	56.824	270.000
637.651	10082.234	38.000	41.772	56.812	338.000
637.901	10089.241	38.000	41.770	56.800	290.000
638.151	10096.284	38.000	41.769	56.788	350.000
638.401	10103.362	38.000	41.767	56.776	451.000
638.651	10110.476	38.000	41.765	56.765	353.000
638.901	10117.626	38.000	41.763	56.753	339.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
639.401	10132.036	38.000	41.760	56.729	315.000
639.651	10139.296	38.000	41.758	56.718	197.000
639.901	10146.593	38.000	41.756	56.706	336.000
640.151	10153.928	38.000	41.755	56.694	235.000
640.401	10161.301	38.000	41.753	56.682	230.000
640.651	10168.712	38.000	41.751	56.671	307.000
640.901	10176.161	38.000	41.749	56.659	335.000
641.151	10183.649	38.000	41.748	56.647	212.000
641.401	10191.176	38.000	41.746	56.635	234.000
641.651	10198.742	38.000	41.744	56.624	208.000
641.901	10206.348	38.000	41.742	56.612	182.000
642.151	10213.994	38.000	41.741	56.600	82.000
642.401	10221.679	38.000	41.739	56.588	163.000
642.651	10229.406	38.000	41.737	56.576	157.000
642.901	10237.172	38.000	41.735	56.565	228.000
643.151	10244.980	38.000	41.734	56.553	200.000
643.401	10252.829	38.000	41.732	56.541	168.000
643.651	10260.720	38.000	41.730	56.529	190.000
643.901	10268.653	38.000	41.728	56.518	213.000
644.151	10276.628	38.000	41.727	56.506	127.000
644.401	10284.645	38.000	41.725	56.494	133.000
644.651	10292.705	38.000	41.723	56.482	112.000
644.901	10300.808	38.000	41.721	56.471	295.000
645.151	10308.956	38.000	41.720	56.459	186.000
645.401	10317.150	38.000	41.718	56.447	56.000
645.651	10325.391	38.000	41.716	56.435	113.000
645.901	10333.680	38.000	41.714	56.424	123.000
646.151	10342.019	38.000	41.713	56.412	202.000
646.401	10350.407	38.000	41.711	56.400	196.000
646.651	10358.848	38.000	41.709	56.388	367.000
646.901	10367.341	38.000	41.707	56.376	341.000
647.151	10375.888	38.000	41.706	56.365	277.000
647.401	10384.489	38.000	41.704	56.353	359.000
647.651	10393.147	38.000	41.702	56.341	385.000
647.901	10401.862	38.000	41.700	56.329	335.000
648.151	10410.636	38.000	41.699	56.318	343.000
648.401	10419.469	38.000	41.697	56.306	365.000
648.651	10428.363	38.000	41.695	56.294	324.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
649.151	10446.338	38.000	41.692	56.271	333.000
649.401	10455.421	38.000	41.690	56.259	280.000
649.651	10464.569	38.000	41.688	56.247	225.000
649.901	10473.783	38.000	41.686	56.235	698.000
650.151	10483.066	38.000	41.685	56.224	539.000
650.401	10492.417	38.000	41.683	56.212	238.000
650.651	10501.837	38.000	41.681	56.200	366.000
650.901	10511.329	38.000	41.680	56.188	223.000
651.151	10520.893	38.000	41.678	56.176	216.000
651.401	10530.530	38.000	41.676	56.165	313.000
651.651	10540.242	38.000	41.674	56.153	298.000
651.901	10550.029	38.000	41.673	56.141	274.000
652.151	10559.893	38.000	41.671	56.129	272.000
652.401	10569.835	38.000	41.669	56.118	373.000
652.651	10579.855	38.000	41.667	56.106	312.000
652.901	10589.953	38.000	41.666	56.094	257.000
653.151	10600.128	38.000	41.664	56.082	406.000
653.401	10610.381	38.000	41.662	56.071	411.000
653.651	10620.710	38.000	41.660	56.059	513.000
653.901	10631.116	38.000	41.659	56.047	873.000
654.151	10641.598	38.000	41.657	56.035	1033.000
654.401	10652.156	38.000	41.655	56.024	803.000
654.651	10662.789	38.000	41.653	56.012	434.000
654.901	10673.498	38.000	41.652	56.000	354.000
655.151	10684.281	38.000	41.652	55.988	530.000
655.401	10695.139	38.000	41.650	55.976	702.000
655.651	10706.072	38.000	41.648	55.963	799.000
655.901	10717.078	38.000	41.646	55.951	798.000
656.151	10728.157	38.000	41.644	55.939	757.000
656.401	10739.310	38.000	41.643	55.927	531.000
656.651	10750.536	38.000	41.641	55.915	439.000
656.901	10761.835	38.000	41.639	55.902	432.000
657.151	10773.206	38.000	41.637	55.890	805.000
657.401	10784.649	38.000	41.635	55.878	4130.000
657.651	10796.163	38.000	41.633	55.866	2896.000
657.901	10807.748	38.000	41.632	55.854	3961.000
658.151	10819.404	38.000	41.630	55.841	1954.000
658.401	10831.130	38.000	41.628	55.829	902.000
658.651	10842.924	38.000	41.626	55.817	647.000
658.901	10854.787	38.000	41.624	55.805	348.000
659.151	10866.718	38.000	41.623	55.793	370.000
659.401	10878.715	38.000	41.621	55.780	549.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
659.901	10902.908	38.000	41.617	55.756	430.000
660.151	10915.102	38.000	41.615	55.744	473.000
660.401	10927.361	38.000	41.614	55.732	388.000
660.651	10939.682	38.000	41.612	55.720	325.000
660.901	10952.067	38.000	41.610	55.707	348.000
661.151	10964.514	38.000	41.608	55.695	414.000
661.401	10977.022	38.000	41.606	55.683	714.000
661.651	10989.590	38.000	41.604	55.671	579.000
661.901	11002.219	38.000	41.603	55.659	549.000
662.151	11014.907	38.000	41.601	55.646	433.000
662.401	11027.654	38.000	41.599	55.634	324.000
662.651	11040.459	38.000	41.597	55.622	504.000
662.901	11053.321	38.000	41.595	55.610	495.000
663.151	11066.239	38.000	41.594	55.598	495.000
663.401	11079.214	38.000	41.592	55.585	412.000
663.651	11092.243	38.000	41.590	55.573	478.000
663.901	11105.327	38.000	41.588	55.561	478.000
664.151	11118.465	38.000	41.586	55.549	580.000
664.401	11131.656	38.000	41.585	55.537	509.000
664.651	11144.899	38.000	41.583	55.524	644.000
664.901	11158.194	38.000	41.581	55.512	786.000
665.151	11171.541	38.000	41.579	55.500	743.000
665.401	11184.937	38.000	41.577	55.488	687.000
665.651	11198.383	38.000	41.576	55.476	738.000
665.901	11211.878	38.000	41.574	55.463	770.000
666.151	11225.422	38.000	41.572	55.451	790.000
666.401	11239.013	38.000	41.570	55.439	877.000
666.651	11252.650	38.000	41.568	55.427	526.000
666.901	11266.334	38.000	41.566	55.415	624.000
667.151	11280.064	38.000	41.565	55.402	677.000
667.401	11293.838	38.000	41.563	55.390	753.000
667.651	11307.657	38.000	41.561	55.378	807.000
667.901	11321.519	38.000	41.559	55.366	860.000
668.151	11335.423	38.000	41.557	55.354	865.000
668.401	11349.370	38.000	41.556	55.341	719.000
668.651	11363.358	38.000	41.554	55.329	892.000
668.901	11377.387	38.000	41.552	55.317	1406.000
669.151	11391.455	38.000	41.550	55.305	1698.000
669.401	11405.563	38.000	41.548	55.293	1343.000
669.651	11419.710	38.000	41.547	55.280	1094.000
669.901	11433.894	38.000	41.545	55.268	1405.000
670.151	11448.116	38.000	41.543	55.256	1312.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
670.651	11476.668	38.000	41.539	55.232	1855.000
670.901	11490.997	38.000	41.538	55.220	2215.000
671.151	11505.361	38.000	41.536	55.207	2164.000
671.401	11519.758	38.000	41.534	55.195	2053.000
671.651	11534.188	38.000	41.532	55.183	2077.000
671.901	11548.651	38.000	41.530	55.171	2035.000
672.151	11563.145	38.000	41.528	55.159	2214.000
672.401	11577.671	38.000	41.527	55.146	2209.000
672.651	11592.226	38.000	41.525	55.134	1828.000
672.901	11606.812	38.000	41.523	55.122	2024.000
673.151	11621.426	38.000	41.521	55.110	1877.000
673.401	11636.068	38.000	41.519	55.098	1904.000
673.651	11650.738	38.000	41.518	55.085	2098.000
673.901	11665.435	38.000	41.516	55.073	1923.000
674.151	11680.158	38.000	41.514	55.061	1820.000
674.401	11694.906	38.000	41.512	55.049	1963.000
674.651	11709.680	38.000	41.510	55.037	2064.000
674.901	11724.477	38.000	41.509	55.024	1917.000
675.151	11739.297	38.000	41.507	55.012	2035.000
675.401	11754.141	38.000	41.503	55.000	2029.000
675.651	11769.006	38.000	41.501	54.994	1935.000
675.901	11783.893	38.000	41.500	54.988	1911.000
676.151	11798.800	38.000	41.498	54.982	2210.000
676.401	11813.727	38.000	41.496	54.976	2381.000
676.651	11828.673	38.000	41.494	54.970	2428.000
676.901	11843.638	38.000	41.492	54.963	2083.000
677.151	11858.621	38.000	41.490	54.957	1982.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
677.651	11888.637	38.000	41.487	54.945	2742.000
677.901	11903.669	38.000	41.485	54.939	2533.000
678.151	11918.717	38.000	41.483	54.933	2278.000
678.401	11933.778	38.000	41.481	54.927	2258.000
678.651	11948.854	38.000	41.480	54.921	2301.000
678.901	11963.942	38.000	41.478	54.915	2192.000
679.151	11979.043	38.000	41.476	54.909	1824.000
679.401	11994.155	38.000	41.474	54.902	1941.000
679.651	12009.278	38.000	41.472	54.896	2038.000
679.901	12024.412	38.000	41.471	54.890	1901.000
680.151	12039.555	38.000	41.469	54.884	2216.000
680.401	12054.707	38.000	41.467	54.878	1894.000
680.651	12069.867	38.000	41.465	54.872	1165.000
680.901	12085.034	38.000	41.463	54.866	1470.000
681.151	12100.209	38.000	41.461	54.860	1371.000
681.401	12115.389	38.000	41.460	54.854	1617.000
681.651	12130.575	38.000	41.458	54.848	1507.000
681.901	12145.765	38.000	41.456	54.841	1623.000
682.151	12160.960	38.000	41.454	54.835	1603.000
682.401	12176.158	38.000	41.452	54.829	1951.000
682.651	12191.358	38.000	41.451	54.823	2012.000
682.901	12206.561	38.000	41.449	54.817	2491.000
683.151	12221.765	38.000	41.447	54.811	2510.000
683.401	12236.969	38.000	41.445	54.805	2623.000
683.651	12252.173	38.000	41.443	54.799	2408.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
684.151	12282.578	38.000	41.440	54.787	2696.000
684.401	12297.778	38.000	41.438	54.780	2599.000
684.651	12312.975	38.000	41.436	54.774	2527.000
684.901	12328.168	38.000	41.434	54.768	2739.000
685.151	12343.357	38.000	41.433	54.762	2936.000
685.401	12358.541	38.000	41.431	54.756	2504.000
685.651	12373.719	38.000	41.429	54.750	2611.000
685.901	12388.891	38.000	41.427	54.744	2621.000
686.151	12404.056	38.000	41.425	54.738	2602.000
686.401	12419.213	38.000	41.423	54.732	2558.000
686.651	12434.362	38.000	41.422	54.726	2467.000
686.901	12449.502	38.000	41.420	54.720	2322.000
687.151	12464.632	38.000	41.418	54.713	2195.000
687.401	12479.751	38.000	41.416	54.707	2486.000
687.651	12494.859	38.000	41.414	54.701	1956.000
687.901	12509.956	38.000	41.413	54.695	2029.000
688.151	12525.039	38.000	41.411	54.689	1831.000
688.401	12540.110	38.000	41.409	54.683	1705.000
688.651	12555.166	38.000	41.407	54.677	1518.000
688.901	12570.208	38.000	41.405	54.671	983.000
689.151	12585.235	38.000	41.404	54.665	887.000
689.401	12600.245	38.000	41.402	54.659	922.000
689.651	12615.236	38.000	41.400	54.652	1261.000
689.901	12630.208	38.000	41.398	54.646	1101.000
690.151	12645.157	38.000	41.396	54.640	1543.000
690.401	12660.083	38.000	41.395	54.634	1901.000
690.651	12674.984	38.000	41.393	54.628	1954.000
690.901	12689.857	38.000	41.391	54.622	1624.000
691.151	12704.702	38.000	41.389	54.616	1835.000
691.401	12719.517	38.000	41.387	54.610	1766.000
691.651	12734.299	38.000	41.385	54.604	1658.000
691.901	12749.047	38.000	41.384	54.598	1460.000
692.151	12763.760	38.000	41.382	54.591	1419.000
692.401	12778.435	38.000	41.380	54.585	1183.000
692.651	12793.071	38.000	41.378	54.579	1107.000
692.901	12807.666	38.000	41.376	54.573	888.000
693.151	12822.218	38.000	41.375	54.567	851.000
693.401	12836.727	38.000	41.373	54.561	934.000
693.651	12851.189	38.000	41.371	54.555	834.000
693.901	12865.603	38.000	41.369	54.549	879.000
694.151	12879.968	38.000	41.367	54.543	792.000
694.401	12894.282	38.000	41.366	54.537	814.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
694.901	12922.749	38.000	41.362	54.524	787.000
695.151	12936.898	38.000	41.360	54.518	499.000
695.401	12950.990	38.000	41.358	54.512	616.000
695.651	12965.021	38.000	41.357	54.506	861.000
695.901	12978.991	38.000	41.355	54.500	955.000
696.151	12992.898	38.000	41.353	54.488	784.000
696.401	13006.739	38.000	41.351	54.476	667.000
696.651	13020.514	38.000	41.349	54.463	716.000
696.901	13034.220	38.000	41.347	54.451	844.000
697.151	13047.856	38.000	41.346	54.439	908.000
697.401	13061.420	38.000	41.344	54.427	752.000
697.651	13074.910	38.000	41.342	54.415	938.000
697.901	13088.325	38.000	41.340	54.402	1146.000
698.151	13101.664	38.000	41.338	54.390	712.000
698.401	13114.927	38.000	41.337	54.378	805.000
698.651	13128.115	38.000	41.335	54.366	880.000
698.901	13141.227	38.000	41.333	54.354	941.000
699.151	13154.264	38.000	41.331	54.341	1417.000
699.401	13167.226	38.000	41.329	54.329	2360.000
699.651	13180.114	38.000	41.328	54.317	2467.000
699.901	13192.927	38.000	41.326	54.305	1874.000
700.151	13205.665	38.000	41.324	54.293	1042.000
700.401	13218.330	38.000	41.322	54.280	1015.000
700.651	13230.921	38.000	41.320	54.268	1604.000
700.901	13243.438	38.000	41.318	54.256	1738.000
701.151	13255.881	38.000	41.317	54.244	1085.000
701.401	13268.252	38.000	41.315	54.232	1025.000
701.651	13280.549	38.000	41.313	54.220	1249.000
701.901	13292.774	38.000	41.311	54.207	1570.000
702.151	13304.926	38.000	41.309	54.195	1857.000
702.401	13317.006	38.000	41.308	54.183	1784.000
702.651	13329.014	38.000	41.306	54.171	1709.000
702.901	13340.949	38.000	41.304	54.159	1471.000
703.151	13352.813	38.000	41.302	54.146	1499.000
703.401	13364.606	38.000	41.300	54.134	1619.000
703.651	13376.327	38.000	41.299	54.122	1461.000
703.901	13387.978	38.000	41.297	54.110	1480.000
704.151	13399.557	38.000	41.295	54.098	1300.000
704.401	13411.066	38.000	41.293	54.085	1173.000
704.651	13422.504	38.000	41.291	54.073	1100.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
705.151	13445.171	38.000	41.288	54.049	1041.000
705.401	13456.399	38.000	41.286	54.037	788.000
705.651	13467.558	38.000	41.284	54.024	675.000
705.901	13478.647	38.000	41.282	54.012	676.000
706.151	13489.668	38.000	41.280	54.000	797.000
706.401	13500.619	38.000	41.279	53.988	823.000
706.651	13511.502	38.000	41.277	53.976	631.000
706.901	13522.317	38.000	41.275	53.963	452.000
707.151	13533.063	38.000	41.273	53.951	552.000
707.401	13543.741	38.000	41.271	53.939	816.000
707.651	13554.351	38.000	41.270	53.927	717.000
707.901	13564.894	38.000	41.268	53.915	695.000
708.151	13575.369	38.000	41.266	53.902	1120.000
708.401	13585.777	38.000	41.264	53.890	1381.000
708.651	13596.118	38.000	41.262	53.878	1640.000
708.901	13606.393	38.000	41.261	53.866	1386.000
709.151	13616.600	38.000	41.259	53.854	931.000
709.401	13626.742	38.000	41.257	53.841	870.000
709.651	13636.817	38.000	41.255	53.829	606.000
709.901	13646.827	38.000	41.253	53.817	597.000
710.151	13656.770	38.000	41.252	53.805	744.000
710.401	13666.649	38.000	41.250	53.793	867.000
710.651	13676.462	38.000	41.248	53.780	1313.000
710.901	13686.210	38.000	41.246	53.768	903.000
711.151	13695.893	38.000	41.244	53.756	896.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
711.651	13715.066	38.000	41.241	53.732	976.000
711.901	13724.556	38.000	41.239	53.720	790.000
712.151	13733.982	38.000	41.237	53.707	661.000
712.401	13743.345	38.000	41.235	53.695	735.000
712.651	13752.644	38.000	41.233	53.683	870.000
712.901	13761.879	38.000	41.232	53.671	808.000
713.151	13771.052	38.000	41.230	53.659	688.000
713.401	13780.161	38.000	41.228	53.646	1615.000
713.651	13789.208	38.000	41.226	53.634	2199.000
713.901	13798.192	38.000	41.224	53.622	2174.000
714.151	13807.115	38.000	41.223	53.610	2154.000
714.401	13815.975	38.000	41.221	53.598	2119.000
714.651	13824.773	38.000	41.219	53.585	2180.000
714.901	13833.510	38.000	41.217	53.573	1781.000
715.151	13842.185	38.000	41.215	53.561	2141.000
715.401	13850.800	38.000	41.213	53.549	2173.000
715.651	13859.353	38.000	41.212	53.537	1783.000
715.901	13867.846	38.000	41.210	53.524	1854.000
716.151	13876.278	38.000	41.208	53.512	1515.000
716.401	13884.649	38.000	41.206	53.500	1252.000
716.651	13892.961	38.000	41.205	53.495	1518.000
716.901	13901.213	38.000	41.203	53.489	2021.000
717.151	13909.405	38.000	41.202	53.484	2336.000
717.401	13917.538	38.000	41.200	53.479	2771.000
717.651	13925.611	38.000	41.198	53.473	3118.000
717.901	13933.626	38.000	41.197	53.468	2973.000
718.151	13941.581	38.000	41.195	53.463	3485.000
718.401	13949.478	38.000	41.194	53.457	3596.000
718.651	13957.317	38.000	41.192	53.452	3843.000
718.901	13965.098	38.000	41.190	53.447	4316.000
719.151	13972.820	38.000	41.189	53.441	4582.000
719.401	13980.485	38.000	41.187	53.436	4938.000
719.651	13988.093	38.000	41.186	53.431	4988.000
719.901	13995.643	38.000	41.184	53.425	4835.000
720.151	14003.136	38.000	41.183	53.420	4730.000
720.401	14010.572	38.000	41.181	53.415	4749.000
720.651	14017.952	38.000	41.179	53.409	4841.000
720.901	14025.275	38.000	41.178	53.404	4963.000
721.151	14032.542	38.000	41.176	53.399	4878.000
721.401	14039.753	38.000	41.175	53.393	5092.000
721.651	14046.908	38.000	41.173	53.388	4907.000
721.901	14054.007	38.000	41.171	53.383	4800.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
722.401	14068.041	38.000	41.168	53.372	4988.000
722.651	14074.975	38.000	41.167	53.367	4766.000
722.901	14081.854	38.000	41.165	53.361	4702.000
723.151	14088.679	38.000	41.164	53.356	4615.000
723.401	14095.450	38.000	41.162	53.351	4603.000
723.651	14102.166	38.000	41.160	53.345	4434.000
723.901	14108.829	38.000	41.159	53.340	4580.000
724.151	14115.438	38.000	41.157	53.335	2209.000
724.401	14121.994	38.000	41.156	53.329	33.000
727.601	14201.282	38.000	41.135	53.261	1802.000
727.851	14207.123	38.000	41.134	53.256	1887.000
728.101	14212.913	38.000	41.132	53.251	2207.000
728.351	14218.653	38.000	41.131	53.245	2140.000
728.601	14224.344	38.000	41.129	53.240	2808.000
728.851	14229.984	38.000	41.127	53.235	3511.000
729.101	14235.576	38.000	41.126	53.229	3651.000
729.351	14241.118	38.000	41.124	53.224	3540.000
729.601	14246.611	38.000	41.123	53.219	3267.000
729.851	14252.056	38.000	41.121	53.213	2492.000
730.101	14257.452	38.000	41.120	53.208	2211.000
730.351	14262.800	38.000	41.118	53.203	1974.000
730.601	14268.099	38.000	41.116	53.197	1485.000
730.851	14273.351	38.000	41.115	53.192	1529.000
731.101	14278.555	38.000	41.113	53.187	1096.000
731.351	14283.712	38.000	41.112	53.181	1095.000
731.601	14288.822	38.000	41.110	53.176	1194.000
731.851	14293.884	38.000	41.108	53.171	1080.000
732.101	14298.900	38.000	41.107	53.165	1337.000
732.351	14303.869	38.000	41.105	53.160	1383.000
732.601	14308.792	38.000	41.104	53.155	1294.000
732.851	14313.668	38.000	41.102	53.149	1015.000
733.101	14318.499	38.000	41.101	53.144	875.000
733.351	14323.284	38.000	41.099	53.139	652.000
733.601	14328.023	38.000	41.097	53.133	618.000
733.851	14332.717	38.000	41.096	53.128	555.000
734.101	14337.367	38.000	41.094	53.123	528.000
734.351	14341.971	38.000	41.093	53.117	4085.000
734.601	14346.530	38.000	41.091	53.112	5113.000
734.851	14351.045	38.000	41.089	53.107	5337.000
735.101	14355.516	38.000	41.088	53.101	4432.000
735.351	14359.943	38.000	41.086	53.096	4317.000
735.601	14364.326	38.000	41.085	53.091	4449.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
736.101	14372.962	38.000	41.082	53.080	4580.000
736.351	14377.215	38.000	41.080	53.075	4530.000
736.601	14381.425	38.000	41.078	53.069	3989.000
736.851	14385.593	38.000	41.077	53.064	3774.000
737.101	14389.718	38.000	41.075	53.059	4020.000
737.351	14393.800	38.000	41.074	53.053	3773.000
737.601	14397.841	38.000	41.072	53.048	3518.000
737.851	14401.840	38.000	41.070	53.043	4433.000
738.101	14405.797	38.000	41.069	53.037	4479.000
738.351	14409.713	38.000	41.067	53.032	3822.000
738.601	14413.587	38.000	41.066	53.027	2935.000
738.851	14417.421	38.000	41.064	53.021	4295.000
739.101	14421.213	38.000	41.063	53.016	4841.000
739.351	14424.966	38.000	41.061	53.011	5112.000
739.601	14428.677	38.000	41.059	53.005	5023.000
739.851	14432.349	38.000	41.058	53.000	5255.000
740.101	14435.981	38.000	41.056	52.991	4971.000
740.351	14439.573	38.000	41.054	52.983	4454.000
740.601	14443.126	38.000	41.052	52.974	4004.000
740.851	14446.639	38.000	41.050	52.965	4004.000
741.101	14450.113	38.000	41.049	52.957	3696.000
741.351	14453.549	38.000	41.047	52.948	3485.000
741.601	14456.945	38.000	41.045	52.940	4041.000
741.851	14460.304	38.000	41.043	52.931	4226.000
742.101	14463.624	38.000	41.041	52.922	4401.000
742.351	14466.906	38.000	41.039	52.914	4023.000
742.601	14470.151	38.000	41.038	52.905	4596.000
742.851	14473.358	38.000	41.036	52.896	4909.000
743.101	14476.527	38.000	41.034	52.888	5238.000
743.351	14479.660	38.000	41.032	52.879	5759.000
743.601	14482.755	38.000	41.030	52.870	5435.000
743.851	14485.814	38.000	41.028	52.862	5477.000
744.101	14488.837	38.000	41.027	52.853	5355.000
744.351	14491.823	38.000	41.025	52.844	5409.000
744.601	14494.773	38.000	41.023	52.836	5493.000
744.851	14497.687	38.000	41.021	52.827	5471.000
745.101	14500.566	38.000	41.019	52.819	5351.000
745.351	14503.409	38.000	41.017	52.810	5558.000
745.601	14506.217	38.000	41.016	52.801	5620.000
745.851	14508.990	38.000	41.014	52.793	5698.000
746.101	14511.729	38.000	41.012	52.784	5722.000
746.351	14514.433	38.000	41.010	52.775	5572.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
746.851	14519.738	38.000	41.006	52.758	5825.000
747.101	14522.339	38.000	41.005	52.749	5790.000
747.351	14524.907	38.000	41.003	52.741	5779.000
747.601	14527.441	38.000	41.001	52.732	5973.000
747.851	14529.942	38.000	40.999	52.723	6239.000
748.101	14532.411	38.000	40.997	52.715	6244.000
748.351	14534.846	38.000	40.996	52.706	6320.000
748.601	14537.249	38.000	40.994	52.698	6380.000
748.851	14539.619	38.000	40.992	52.689	6219.000
749.101	14541.957	38.000	40.990	52.680	6587.000
749.351	14544.263	38.000	40.988	52.672	6666.000
749.601	14546.538	38.000	40.986	52.663	6910.000
749.851	14548.780	38.000	40.985	52.654	6828.000
750.101	14550.992	38.000	40.983	52.646	6746.000
750.351	14553.173	38.000	40.981	52.637	6724.000
750.601	14555.322	38.000	40.979	52.628	5467.000
750.851	14557.441	38.000	40.977	52.620	5902.000
751.101	14559.530	38.000	40.975	52.611	6130.000
751.351	14561.588	38.000	40.974	52.602	6292.000
751.601	14563.616	38.000	40.972	52.594	5799.000
751.851	14565.614	38.000	40.970	52.585	5940.000
752.101	14567.583	38.000	40.968	52.577	5484.000
752.351	14569.523	38.000	40.966	52.568	5658.000
752.601	14571.433	38.000	40.964	52.559	5875.000
752.851	14573.314	38.000	40.963	52.551	5677.000
753.101	14575.167	38.000	40.961	52.542	6061.000
753.351	14576.991	38.000	40.959	52.533	5862.000
753.601	14578.787	38.000	40.957	52.525	6074.000
753.851	14580.555	38.000	40.955	52.516	6155.000
754.101	14582.295	38.000	40.953	52.507	6082.000
754.351	14584.007	38.000	40.952	52.499	6586.000
754.601	14585.692	38.000	40.950	52.490	6772.000
754.851	14587.349	38.000	40.948	52.481	6864.000
755.101	14588.980	38.000	40.946	52.473	7020.000
755.351	14590.583	38.000	40.944	52.464	6799.000
755.601	14592.160	38.000	40.942	52.456	5974.000
755.851	14593.711	38.000	40.941	52.447	5671.000
756.101	14595.236	38.000	40.939	52.438	5242.000
756.351	14596.735	38.000	40.937	52.430	5490.000
756.601	14598.208	38.000	40.935	52.421	5039.000
756.851	14599.655	38.000	40.933	52.412	5134.000
757.101	14601.077	38.000	40.931	52.404	4848.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
757.601	14603.847	38.000	40.928	52.386	4608.000
757.851	14605.194	38.000	40.926	52.378	4829.000
758.101	14606.517	38.000	40.924	52.369	4999.000
758.351	14607.816	38.000	40.922	52.360	5021.000
758.601	14609.091	38.000	40.920	52.352	5084.000
758.851	14610.342	38.000	40.919	52.343	5021.000
759.101	14611.570	38.000	40.917	52.335	4824.000
759.351	14612.774	38.000	40.915	52.326	4793.000
759.601	14613.955	38.000	40.913	52.317	4351.000
759.851	14615.114	38.000	40.911	52.309	4624.000
760.101	14616.249	38.000	40.909	52.300	4555.000
760.351	14617.362	38.000	40.909	52.294	4537.000
760.601	14618.453	38.000	40.908	52.288	5113.000
760.851	14619.521	38.000	40.906	52.282	5286.000
761.101	14620.568	38.000	40.904	52.276	5251.000
761.351	14621.593	38.000	40.902	52.270	4888.000
761.601	14622.597	38.000	40.900	52.263	3793.000
761.851	14623.580	38.000	40.899	52.257	4349.000
762.101	14624.541	38.000	40.897	52.251	4346.000
762.351	14625.482	38.000	40.895	52.245	4077.000
762.601	14626.402	38.000	40.893	52.239	4279.000
762.851	14627.302	38.000	40.891	52.233	4337.000
763.101	14628.182	38.000	40.889	52.227	4284.000
763.351	14629.042	38.000	40.888	52.221	4097.000
763.601	14629.882	38.000	40.886	52.215	4137.000
763.851	14630.702	38.000	40.884	52.209	4299.000
764.101	14631.504	38.000	40.882	52.202	4012.000
764.351	14632.286	38.000	40.880	52.196	4838.000
764.601	14633.050	38.000	40.879	52.190	5102.000
764.851	14633.795	38.000	40.877	52.184	5013.000
765.101	14634.521	38.000	40.875	52.178	5039.000
765.351	14635.229	38.000	40.873	52.172	5508.000
765.601	14635.920	38.000	40.871	52.166	5563.000
765.851	14636.592	38.000	40.870	52.160	5562.000
766.101	14637.247	38.000	40.868	52.154	5469.000
766.351	14637.885	38.000	40.866	52.148	5451.000
766.601	14638.505	38.000	40.864	52.141	5619.000
766.851	14639.109	38.000	40.862	52.135	5441.000
767.101	14639.696	38.000	40.861	52.129	5749.000
767.351	14640.266	38.000	40.859	52.123	5501.000
767.601	14640.821	38.000	40.857	52.117	5914.000
767.851	14641.359	38.000	40.855	52.111	5308.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
768.351	14642.388	38.000	40.851	52.099	5586.000
768.601	14642.879	38.000	40.850	52.093	5756.000
768.851	14643.355	38.000	40.848	52.087	5535.000
769.101	14643.816	38.000	40.846	52.080	5786.000
769.351	14644.263	38.000	40.844	52.074	5615.000
769.601	14644.695	38.000	40.842	52.068	5840.000
769.851	14645.112	38.000	40.841	52.062	6163.000
770.101	14645.516	38.000	40.839	52.056	6034.000
770.351	14645.905	38.000	40.837	52.050	5786.000
770.601	14646.281	38.000	40.835	52.044	5868.000
770.851	14646.644	38.000	40.833	52.038	5944.000
771.101	14646.993	38.000	40.832	52.032	6117.000
771.351	14647.329	38.000	40.830	52.026	5997.000
771.601	14647.652	38.000	40.828	52.020	6163.000
771.851	14647.963	38.000	40.826	52.013	5907.000
772.101	14648.261	38.000	40.824	52.007	5711.000
772.351	14648.548	38.000	40.822	52.001	5984.000
772.601	14648.822	38.000	40.821	51.995	6280.000
772.851	14649.084	38.000	40.819	51.989	5886.000
773.101	14649.335	38.000	40.817	51.983	6307.000
773.351	14649.575	38.000	40.815	51.977	5781.000
773.601	14649.803	38.000	40.813	51.971	5837.000
773.851	14650.021	38.000	40.812	51.965	6377.000
774.101	14650.228	38.000	40.810	51.959	5789.000
774.351	14650.425	38.000	40.808	51.952	5294.000
774.601	14650.611	38.000	40.806	51.946	5046.000
774.851	14650.787	38.000	40.804	51.940	4835.000
775.101	14650.954	38.000	40.803	51.934	5312.000
775.351	14651.111	38.000	40.801	51.928	6341.000
775.601	14651.258	38.000	40.799	51.922	6016.000
775.851	14651.397	38.000	40.797	51.916	5999.000
776.101	14651.526	38.000	40.795	51.910	6220.000
776.351	14651.647	38.000	40.794	51.904	5959.000
776.601	14651.759	38.000	40.792	51.898	5876.000
776.851	14651.863	38.000	40.790	51.891	6007.000
777.101	14651.959	38.000	40.788	51.885	5824.000
777.351	14652.046	38.000	40.786	51.879	5978.000
777.601	14652.127	38.000	40.784	51.873	5894.000
777.851	14652.200	38.000	40.783	51.867	6238.000
778.101	14652.265	38.000	40.781	51.861	6217.000
778.351	14652.324	38.000	40.779	51.855	6283.000
778.601	14652.375	38.000	40.777	51.849	6132.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
779.101	14652.460	38.000	40.774	51.837	6096.000
779.351	14652.492	38.000	40.772	51.830	5962.000
779.601	14652.519	38.000	40.770	51.824	6048.000
779.851	14652.540	38.000	40.768	51.818	6116.000
780.101	14652.556	38.000	40.766	51.812	5963.000
780.351	14652.566	38.000	40.765	51.806	6588.000
780.601	14652.571	38.000	40.761	51.800	5572.000
780.851	14652.571	38.000	40.759	51.796	4171.000
781.101	14652.567	38.000	40.757	51.793	3770.000
781.351	14652.558	38.000	40.756	51.789	8309.000
781.601	14652.545	38.000	40.754	51.785	4876.000
781.851	14652.528	38.000	40.752	51.782	3442.000
782.101	14652.508	38.000	40.750	51.778	3565.000
782.351	14652.483	38.000	40.748	51.774	3876.000
782.601	14652.455	38.000	40.746	51.771	4403.000
782.851	14652.425	38.000	40.745	51.767	3869.000
783.101	14652.391	38.000	40.743	51.763	2886.000
783.351	14652.354	38.000	40.741	51.760	5136.000
783.601	14652.316	38.000	40.739	51.756	5435.000
783.851	14652.274	38.000	40.737	51.752	4861.000
784.101	14652.231	38.000	40.736	51.749	5532.000
784.351	14652.186	38.000	40.734	51.745	5686.000
784.601	14652.139	38.000	40.732	51.741	5828.000
784.851	14652.091	38.000	40.730	51.738	7131.000
785.101	14652.042	38.000	40.728	51.734	6505.000
785.351	14651.991	38.000	40.727	51.730	6024.000
785.601	14651.940	38.000	40.725	51.727	5532.000
785.851	14651.888	38.000	40.723	51.723	4977.000
786.101	14651.836	38.000	40.721	51.720	4681.000
786.351	14651.784	38.000	40.719	51.716	4106.000
786.601	14651.732	38.000	40.717	51.712	4993.000
786.851	14651.680	38.000	40.716	51.709	4801.000
787.101	14651.629	38.000	40.714	51.705	4736.000
787.351	14651.578	38.000	40.712	51.701	4914.000
787.601	14651.529	38.000	40.710	51.698	6683.000
787.851	14651.480	38.000	40.708	51.694	4395.000
788.101	14651.433	38.000	40.707	51.690	4920.000
788.351	14651.388	38.000	40.705	51.687	3993.000
788.601	14651.344	38.000	40.703	51.683	3423.000
788.851	14651.302	38.000	40.701	51.679	4211.000
789.101	14651.263	38.000	40.699	51.676	5270.000
789.351	14651.226	38.000	40.698	51.672	4844.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
789.851	14651.160	38.000	40.694	51.665	4491.000
790.101	14651.132	38.000	40.692	51.661	3834.000
790.351	14651.107	38.000	40.690	51.657	4045.000
790.601	14651.086	38.000	40.689	51.654	4680.000
790.851	14651.068	38.000	40.687	51.650	4997.000
791.101	14651.054	38.000	40.685	51.646	5331.000
791.351	14651.045	38.000	40.683	51.643	6226.000
791.601	14651.039	38.000	40.681	51.639	5221.000
791.851	14651.039	38.000	40.679	51.635	4724.000
792.101	14651.043	38.000	40.678	51.632	4031.000
792.351	14651.052	38.000	40.676	51.628	4573.000
792.601	14651.067	38.000	40.674	51.624	5279.000
792.851	14651.087	38.000	40.672	51.621	5174.000
793.101	14651.112	38.000	40.670	51.617	5652.000
793.351	14651.144	38.000	40.669	51.613	5493.000
793.601	14651.182	38.000	40.667	51.610	5406.000
793.851	14651.226	38.000	40.665	51.606	5971.000
794.101	14651.276	38.000	40.663	51.602	6089.000
794.351	14651.333	38.000	40.661	51.599	4986.000
794.601	14651.398	38.000	40.660	51.595	4355.000
794.851	14651.469	38.000	40.658	51.591	5211.000
795.101	14651.548	38.000	40.656	51.588	5886.000
795.351	14651.634	38.000	40.654	51.584	5406.000
795.601	14651.729	38.000	40.652	51.580	5742.000
795.851	14651.831	38.000	40.651	51.577	5837.000
796.101	14651.941	38.000	40.649	51.573	5916.000
796.351	14652.060	38.000	40.647	51.570	5967.000
796.601	14652.188	38.000	40.645	51.566	6545.000
796.851	14652.325	38.000	40.643	51.562	7343.000
797.101	14652.470	38.000	40.641	51.559	6368.000
797.351	14652.626	38.000	40.640	51.555	6381.000
797.601	14652.790	38.000	40.638	51.551	6256.000
797.851	14652.965	38.000	40.636	51.548	5693.000
798.101	14653.149	38.000	40.634	51.544	4676.000
798.351	14653.344	38.000	40.632	51.540	3984.000
798.601	14653.549	38.000	40.631	51.537	5670.000
798.851	14653.764	38.000	40.629	51.533	5454.000
799.101	14653.991	38.000	40.627	51.529	4873.000
799.351	14654.228	38.000	40.625	51.526	5508.000
799.601	14654.477	38.000	40.623	51.522	6893.000
799.851	14654.737	38.000	40.622	51.518	6295.000
800.101	14655.009	38.000	40.620	51.515	6194.000

Depth (cm)	Age (yrs)	Modern Precip. (cm)	Precip. (10% Wetter) (cm)	Precip. (50% Wetter) (cm)	Ti (counts)
800.601	14655.589	38.000	40.616	51.507	4339.000
800.851	14655.897	38.000	40.614	51.504	44.000
801.101	14656.218	38.000	40.613	51.500	27.000

APPENDIX F

CARBONATE PERCENTAGES

Equation 4 in Chapter 4, Appendix 1 estimates terrigenous input as a percentage of sediment in the core. Figure 4.12 illustrates the model outputs if CaCO_3 and total carbon are used in this estimate (where $E/S = 1 - \% \text{Carbonate or total carbon}$). We measured percent carbonate in our lab via carbonate digestion and titration. Krause and Whitlock (2013) report CaCO_3 and total carbon values. Both sets of data are reported below.

Dixon Lab Measured CaCO₃

Sample	Depth (cm)	CaCO₃ (weight %)
1	544.23	75.7
2	584.85	86.1
3	615.85	73.4
4	633.79	95.7
5	651.90	97.7
6	665.03	78.2
7	678.03	58.4
8	691.90	72.4
9	705.03	45
10	720.53	14.8
11	748.85	45.4
12	748.35	10.6
13	764.98	8.9
14	776.98	0.88
15	784.98	8.9
16	801.57	7.9

Krause and Whitlock (2013)

Depth (cm)	Total Carbon (weight %)	CaCO₃ (weight %)
525.80	40.08	2.60
529.10	24.59	17.78
532.10	17.34	2.36
535.30	25.69	13.48
538.60	5.36	1.67
541.70	30.70	16.65
544.65	21.03	22.99
547.75	28.59	18.34
550.80	25.43	22.18
#N/A	21.34	22.80
556.90	21.37	24.33
560.10	23.31	24.94
563.60	41.60	12.64
567.10	18.36	26.65
571.10	12.50	31.29
575.10	17.42	25.36
579.60	14.52	27.63
584.60	19.71	22.14
589.85	13.69	21.91
595.35	15.30	27.80
#N/A	20.86	23.44
607.35	14.73	29.64
613.35	18.92	25.98
619.10	26.27	21.34
624.85	#N/A	#N/A
629.90	25.03	19.86
634.65	22.17	21.22
639.15	10.94	2.87
643.15	23.06	23.50
646.90	35.26	12.14
650.40	26.71	20.90
653.40	33.58	12.31
656.40	26.08	12.20
659.00	44.98	2.72
661.65	19.15	1.86

#N/A	48.77	3.00
666.40	53.27	2.80
668.65	45.36	2.84
670.90	48.40	2.78
673.00	39.84	2.49
675.15	11.97	1.22
677.20	3.32	1.11
678.40	2.94	1.00
681.40	2.47	1.17
683.50	2.85	1.07
685.50	3.33	2.42
687.65	2.76	1.12
689.65	2.18	1.02
691.85	1.82	0.98
693.90	2.02	1.09
696.15	2.37	1.12
698.40	1.75	1.23
700.65	2.09	1.40
703.15	2.14	1.19
705.90	1.60	1.48
708.65	1.88	1.11
711.65	2.07	1.25
714.90	1.84	1.19
718.40	1.82	1.32
722.40	1.97	1.07
727.60	1.65	1.22
731.85	1.92	1.65
738.30	1.56	1.55