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ICE FLUCTUATIONS IN MORaine canyon, Antarctica DATED WITH
COSmoGENic $^{26}$Al, $^{10}$Be, and $^{21}$Ne

by

Emma Mirabelli Lord
Bachelor of Arts, Green Mountain College, 2013

A Master’s Thesis
Submitted to the Graduate Faculty
of the
University of North Dakota
in partial fulfillment of the requirements
for the degree of
Master of Science

Grand Forks, North Dakota
May
2016
This thesis, submitted by Emma M. Lord in partial fulfillment of the requirements for the Degree of Master of Science from the University of North Dakota, has been read by the Faculty Advisory Committee under whom the work has been done and is hereby approved.

Dr. Jaakko Putkonen, Chairperson

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This thesis is being submitted by the appointed advisory committee as having met all of the requirements of the School of Graduate Studies at the University of North Dakota and is hereby approved.

Dr. Wayne Swisher
Dean of the School of Graduate Studies

April 29, 2016

Date
PERMISSION

Title: Ice Fluctuations in Moraine Canyon, Antarctica Dated with Cosmogenic $^{26}$Al, $^{10}$Be, and $^{21}$Ne

Department: Harold Hamm School of Geology and Geological Engineering

Degree: Master of Science

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Emma M. Lord
4/14/16
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To my family who has tirelessly supported me in all my endeavors
ABSTRACT

Knowledge and understanding of past Antarctic ice sheet behavior is necessary to illustrate ice sheet response to our currently warming climate and the ice sheet’s contribution to global sea level change. Our knowledge of glacial fluctuations in the southern Transantarctic Mountains (TAM) is limited because few studies in this area have been done; therefore much information can be gained from further investigations in this region. Glacial deposits in Moraine Canyon, a tributary of Amundsen Glacier in the southern TAM, show numerous periods of ice thickening. Moraine Canyon responds to changes in the thickness of Amundsen Glacier, and Amundsen Glacier responds to changes in thickness of the East Antarctic and West Antarctic Ice Sheets (EAIS and WAIS). \(^{10}\text{Be}, {^{26}\text{Al}}, \text{and} {^{21}\text{Ne}}\) concentrations from vertical profiles in \(~1\text{m}\) deep soil pits from four glacial deposits provide exposure ages and local regolith erosion rates. Exposure ages from three lateral moraines and one valley floor deposit show an overall decrease in ice levels in Moraine Canyon since at least \(1.21\text{ Ma}\) punctuated by at least four periods of ice thickening of Amundsen Glacier. Local regolith erosion rates range from \(0.21-4.40\text{ m Ma}^{-1}\).

During subsequent periods of ice thickening, lateral moraines were deposited at \(1.21\text{ Ma}, 1.10\text{ Ma}, \text{and} 80.4\text{ ka}\). The regolith overlying glacial ice on the valley floor was deposited \(287\text{ ka}\). A period of significant ice thinning in Moraine Canyon occurred between \(1.10\text{ Ma}\) and \(287\text{ ka}\). During this thinning event, ice levels of Amundsen Glacier were low enough to allow the glacial ice in Moraine Canyon to completely flow out of the valley. Significant thinning of the EAIS at the head of Amundsen Glacier was necessary for this amount of ice lowering to occur during this
time. This research provides new constraints on past fluctuations of the EAIS and suggests that it is more sensitive to shorter term climatic fluctuations than previously thought.
CHAPTER I
GENERAL INTRODUCTION AND BACKGROUND

The changing climate and its effect on the stability of the Antarctic ice sheets are topics of great interest and importance to scientists and policymakers alike. Decreasing volume of the continental scale ice sheets covering most of the Antarctic landmass has worldwide ramifications, particularly the ice sheets’ contributions to global sea level rise. Paleoclimate information deduced from ice cores, offshore sediment cores, and in-situ glacial deposits enable us to reconstruct the timing of past ice sheet fluctuations. This knowledge is necessary to create reliable models that can accurately predict future changes in global sea level.

The lack of information on past ice sheet behavior is an impediment to solving this puzzle. The longest continuous climate record for Antarctica comes from ice cores drilled at Dome C which date back to 800,000 years (EPICA, 2009). In-situ geologic field evidence from various locations throughout the continent allows us to extend Antarctica’s climate record back to several million years.

Making up most of the ice free land in Antarctica, the Transantarctic Mountains (TAM) form a barrier between the East and West Antarctic Ice Sheets (EAIS and WAIS). Ice from the high elevation polar plateau of the EAIS drains down through the TAM via large outlet glaciers including Byrd, Nimrod, Beardmore, Scott, and Amundsen Glaciers in to the Ross Sea and WAIS. As much as 90% of the discharge from the EAIS flows down these rapidly flowing outlet glaciers and ice streams (Bamber et al., 2000). Throughout the TAM, outlet glaciers fluctuate in
thickness in response to changing ice thickness and volume of the EAIS and WAIS. Expansion of the ice sheets is recorded by glacial deposits and exposed bedrock along these outlet glaciers and their tributaries. These glacial deposits act as a gage from which ice level changes can be measured. Being that the vast majority of the Antarctic continent is covered by ice, EAIS outlet glaciers and their associated deposits provide some of the only terrestrial records for ice sheet fluctuations. From these glacial deposits past ice sheet thickness can be inferred and used as a proxy for climatic changes in Antarctica.

It is generally accepted that the marine-based WAIS has fluctuated substantially over the past few million years (e.g. Naish et al., 2009). Currently, air temperatures over the EAIS are not warm enough for significant ice melting to occur, but changes in sea level and ocean-induced melting strongly affect the WAIS (McKay et al., 2012). Ice sheet models show that the WAIS is much more responsive to climate change, with full to partial collapse of the WAIS possibly occurring several times over the past few million years (Pollard and DeConto, 2009). Geologic field evidence paired with our understanding of the mechanisms by which the WAIS evolves is used to determine the timing and extent of the ice sheet’s fluctuations.

Understanding past WAIS behavior will improve our understanding and modeling of future fluctuations in relation to sea level change, particularly because the marine-based WAIS responds rapidly to changing ocean and air temperatures. This is particularly concerning as global air and sea temperatures have been rapidly increasing since pre-industrial times. Fluctuations of the WAIS since the Last Glacial Maximum (LGM) are best understood. During the LGM ice highstand, outlet glaciers throughout the TAM experienced considerably higher, as much as 1,250 m, ice levels near their confluence with the WAIS (e.g. Brockheim et al., 1989b;
EAIS outlet glaciers have been thinning since their last highstand during the LGM. While changes of the WAIS are better understood, the glacial history of the larger EAIS is less certain. There are two contending hypotheses for the history of the EAIS: stable or dynamic. With the “stable hypothesis”, the case is made that the EAIS has remained at its current size since ~15 Ma (e.g. Sugden et al., 1993). The “dynamic hypothesis” maintains that the EAIS was a warmer, wet-based ice sheet that fluctuated both in extent and volume from 15-3 Ma before reaching its current size in the late Pliocene (e.g. Webb et al., 1984). Geologic evidence shows successive thinning of the EAIS since the Pliocene (5-3 Ma), which has been suggested to have been caused by global cooling which resulted in less moisture transport from the Southern Ocean and less precipitation over the continent (Altmaier et al., 2010; Kong et al., 2010). Evidence of changes in thickness and extent of the EAIS from the continental margins and along the TAM provides a larger database and wider perspective from which to view the ice sheet’s behavior as a whole.

Glacial fluctuations are recorded by surficial deposits in this landscape, and these glacial fluctuations are used as proxies to describe paleoclimate. However, continuous degradation of the deposits makes paleoclimate interpretation more complicated than if the deposits remained unaltered. Extensive work in the McMurdo Dry Valleys (MDV) of Antarctica shows that low erosion rates lead to the almost perfect preservation of landforms for hundreds of thousands to millions of years (Summerfield et al., 1991; Putkonen et al., 2008; Morgan et al., 2010a,b). Much of our understanding of surface processes in Antarctica stems from research in this area. Located on the perimeter of the continent and heavily influenced by the ocean, the Dry Valleys are not
the best representation of the continent as a whole. Observation and measurement of current geomorphic processes enables us to better understand past landscape changes.

Due to the remoteness and inaccessibility of the southern TAM, relatively few studies on glacial fluctuations and sediment transport have been conducted in the region. These studies describe glacial chronologies ranging from several thousand to millions of years in the southern TAM (e.g. Ackert et al. 2007, 2011; Todd et al., 2010; Bromley et al., 2012; Mukhopadhyay et al., 2012). Long term exposure ages > 1 Ma and slow regolith erosion rates attest to the persistent cold desert conditions across the TAM since the middle Miocene (Fogwill et al., 2004; Di Nicola et al., 2012). Glacial chronology studies at Reedy and Scott Glacier focus on the extent of ice thickening during the LGM, and these studies serve as the primary constraint of ice sheet change in the southern TAM during and after the LGM (Bromely, et al., 2010; Todd et al., 2010).

Located adjacent to Amundsen Glacier approximately midway between the EAIS and WAIS, Moraine Canyon is a unique location that is influenced by changes in both of the ice sheets. This research using cosmogenic nuclides provides evidence for past ice levels in Moraine Canyon in the southern TAM as well as information on regolith transport in the region where very few measurements currently exist. Knowledge of glacial fluctuations and sediment transport will help us to better understand geomorphic processes and interpret climatic history in the southern TAM.

**Motivation and Research Objective**

The objective of this research is to assign minimum ages to the glacial deposits and quantify regolith erosion rates in Moraine Canyon in the southern TAM. By analyzing cosmogenic $^{26}$Al, $^{10}$Be, and $^{21}$Ne nuclide concentrations at different depths within each soil pit sampled we can reconstruct concentration-depth profiles to identify any vertical mixing of
regolith and better constrain concentrations of inherited nuclides from previous periods of exposure. Measuring multiple nuclides with different half-lives allows us to determine local erosion rates.

Exposure ages are assigned to four glacial deposits in Moraine Canyon, three being lateral moraines deposited on the valley wall, and one being unconsolidated regolith overlying glacial ice on the valley floor. The reported exposure ages are based on the analysis of cosmogenic nuclide concentrations along a vertical profile through the top ~1 m of regolith in each deposit. Local erosion rates are also reported for this location. The calculated rates of landscape change and exposure ages of the deposits help to elucidate the glacial history and landscape evolution of Moraine Canyon, which are used to infer the ice fluctuations of Amundsen Glacier and the WAIS and EAIS themselves.

**Cosmogenic Nuclide Background**

Constantly bombarding the Earth’s surface, galactic cosmic rays (GCR) generate nuclear reactions in mineral grains of rock material near the surface. These reactions result in the formation of in-situ cosmogenic nuclides that can be measured in the minerals. There are three principal mechanisms for nuclide production: spallation of nucleons, thermal neutron capture, and interactions of fast muons (Lal, 1991). Spallation is the dominant nuclide forming mechanism within the top one meter of the Earth’s surface. $^{26}$Al, $^{10}$Be, and $^{21}$Ne are three commonly used nuclides that form from Si and O in quartz grains. They are often used in conjunction with one another because all three nuclides can be measured in isolated quartz grains.

Exposure dating with cosmogenic nuclides is based on known production rates of different nuclides, although production rates vary with latitude, altitude, and geomagnetic field
strength (Lal, 1991). The concentration of nuclides in a material is a function of the production rate, decay rate, and erosion rate of the surface. Given a known nuclide production rate, the concentration of a nuclide indicates the length of time the material has been exposed to cosmic rays at or near the surface (Figure 1). Exposure dating with cosmogenic nuclides is ideal for the TAM because the influx of cosmic rays increases with elevation and latitude, causing increased production of nuclides in these regions.

Figure 1. Nuclide accumulation over time. Theoretical nuclide accumulation over time with steady erosion typical of Antarctic regolith surfaces.

There are several basic assumptions used with cosmogenic nuclide dating: increased surface or near surface exposure duration results in higher in-situ nuclide concentrations within the material; the cosmic ray flux affecting the known production rates has remained constant over the past 10 Ma; there are no inherited nuclides present in the rock surface from a period of prior exposure; the nuclide production rate decreases with depth at a known rate within the material (Gosse and Phillips, 2001). The thickness of rock material that cosmic rays can
penetrate and form cosmogenic nuclides is the attenuation length. The attenuation length varies with latitude and altitude due to the effects of the geomagnetic field, but the general value used for Antarctica is 150 g cm\(^{-2}\) (Gosse and Phillips, 2001).

The use of multiple nuclides with different characteristics enables more information to be determined from a single sample. \(^{21}\)Ne is a stable nuclide that accumulates indefinitely while \(^{26}\)Al and \(^{10}\)Be are radionuclides with known half-lives (0.708 Ma and 1.36 Ma respectively) (Nishiizumi, 2004; Nishiizumi et al. 2007). Stable isotopes are useful in exposure studies because they can be used on an exposure timescale of up to several million years. For radionuclides, secular equilibrium is reached when the production rate of the nuclide equals the decay rate. At this point, the surface has been exposed for a long enough time period that the measured nuclide concentration only provides a minimum exposure age. Once a sample has reached secular equilibrium, the nuclide concentration is only a function of the local erosion rate, and the exposure age provides only a minimum age.

While overall erosion rates can be established using a single radio nuclide, the use of several radionuclides with different decay rates provides considerably more information on the erosional history of a surface (Lal, 1991). The use of multiple nuclides enables one to determine erosion rates and complex exposure histories that involve periods of burial or partial shielding from GCR. Radionuclides \(^{26}\)Al and \(^{10}\)Be have different production and decay rates, and are produced in quartz at the fixed ratio of 6.75:1 (Balco and Schuster, 2009b). Under simple exposure histories the ratio of the two nuclides will decrease in a known manner based on the decay rates and local erosion rate. The ratio of Al:Be can be visualized with \(^{10}\)Be concentration in a steady state erosion plot (Lal, 1991). The duration of exposure and decay rates of the nuclides determines the constant exposure line on the steady state erosion plot (Lal, 1991). As
the surface nears secular equilibrium, the path of the $^{26}\text{Al}:^{10}\text{Be}$ ratio decreases until the production rates equal the decay rates. A surface undergoing constant exposure and steady erosion over time will produce a lower $^{26}\text{Al}:^{10}\text{Be}$ ratio path (Figure 2).

**Figure 2.** Erosion Island Plot of $^{10}\text{Be}$ vs. $^{26}\text{Al}/^{10}\text{Be}$ ratio. The starting $^{26}\text{Al}:^{10}\text{Be}$ ratio has been normalized from the initial production ratio of 6.75 for $^{26}\text{Al}:^{10}\text{Be}$. The bold black lines delineate the erosion island with the upper bound showing the nuclide concentrations in a sample undergoing zero erosion and the lower bound showing a sample experiencing steady erosion. Samples plotting below the erosion island have undergone complex exposure. Samples cannot plot above the given 6.75:1 production ratio. Samples plotting in the forbidden zone most likely have analytical error.

Erosion of the exposed rock surface removes accumulated nuclides, resulting in lower nuclide concentrations than a non-eroded surface. If erosion is not accounted for there will be an underestimation of the actual exposure age. When a surface is temporarily buried or completely shielded from GCR, then nuclide production halts and nuclide decay determines the concentration and ratio of $^{26}\text{Al}$ and $^{10}\text{Be}$. Because the half-life of $^{26}\text{Al}$ is less than that of $^{10}\text{Be}$, the
The concentration of Al will decrease faster than the concentration of Be, decreasing their ratio of $^{26}\text{Al}$ to $^{10}\text{Be}$ during burial (Lal, 1991).

The use of cosmogenic nuclide exposure dating in Antarctica is complicated by the recycling of old regolith into new deposits. Unconsolidated regolith deposits may contain inherited nuclides from previous periods of exposure. The concentrations of cosmogenic nuclides in the glacial drifts in Moraine Canyon are expected to reveal the age of deposition of the landforms. Because the moraines are composed of unconsolidated regolith that originated as bedrock farther up the valley, there are likely high concentrations of inherited nuclides from previous exposure of the material. If the inherited nuclides are not taken into account, the age of the landform may be overestimated. An understanding of the geomorphic processes occurring in the cold, arid Antarctic environment and how these processes change the landscape is necessary for accurate interpretations of past glacial fluctuations.

![Concentration of Nuclides with Depth](image)

Figure 3. Decreasing Nuclide Concentration with Depth. Hypothetical graph showing decreasing nuclide concentration with depth. The decrease in nuclide production with depth is due mainly to decreasing spallation energy.
Thesis Organization

Chapter 2 will focus on the dating of glacial landforms and the interpretation of the glacial history of Moraine Canyon. Concentrations of $^{10}\text{Be}$, $^{26}\text{Al}$, and $^{21}\text{Ne}$ were measured in unconsolidated glacial deposits on the valley walls and valley floor of Moraine Canyon. Analysis of these nuclide concentrations from ~1m soil pits provides both exposure ages and local erosion rates. Chapter 3 provides a more detailed examination of the results and the conclusions that can be drawn from them.
CHAPTER II

ICE FLUCTUATIONS IN MORAYNE CANYON, ANTARCTICA DATED WITH COSMOGENIC $^{26}$Al, $^{10}$Be, and $^{21}$Ne

Introduction

Outlet glaciers of the East Antarctic Ice Sheet (EAIS) flowing through the Transantarctic Mountains (TAM) fluctuate based on ice levels of the EAIS and West Antarctic Ice Sheet (WAIS) (e.g. Bockheim et al., 1989; Denton et al., 1989a,b). Repeated advances of these outlet glaciers result in the deposition of moraines and glacial erratics throughout the TAM, and these deposits are interpreted as direct evidence of past ice thickening. Changes in thickness and volume of the EAIS and WAIS are of high importance, and understanding past Antarctic Ice Sheet behavior and their contribution to global sea level change is necessary to create models that can reliably predict future changes.

Located in the Queen Maud Mountains of the southern TAM, Moraine Canyon is a unique location that is directly affected by ice levels of Amundsen Glacier which in turn is influenced by ice dynamics of both the EAIS and WAIS. Because of the inaccessibility of the southern TAM, there are relatively few measurements of past ice fluctuations in this area. However, several studies in this region describe glacial chronologies spanning from the Last Glacial Maximum (~20 kya) to several million years (e.g. Ackert et al. 2007, 2011; Todd et al., 2010; Bromley et al., 2012; Mukhopadhyay et al., 2012). The majority of the studies in this region focus on changes of the WAIS, while the behavior of the EAIS is less understood.
In order to address questions of past ice levels in the Amundsen Glacier area, we used cosmogenic nuclide exposure dating to determine depositional ages of unconsolidated glacial deposits in Moraine Canyon. We present depositional ages and erosion rates of lateral moraines and supraglacial regolith deposits in Moraine Canyon based on cosmic ray produced $^{10}$Be, $^{26}$Al, and $^{21}$Ne from ~1 m deep soil pits in each deposit. Ice level changes in Moraine Canyon are reconstructed from geologic evidence, and these fluctuations are related to changes in ice thickness of Amundsen Glacier at the mouth of the valley and fluctuations in the ice levels of the WAIS and less understood EAIS.

**Field Area and Sample Sites**

Moraine Canyon (158° 10’ W, 86° 7’ S) is a 25 km long, 4 km wide valley located in the Queen Maud Mountains of the TAM (Figure 4, 5). Situated between Scott and Amundsen Glaciers and adjacent to the Nilsen Plateau, the valley is composed of two main branches, one being exposed glacial ice and the other being dirty debris covered glacial ice. Numerous distinct lateral moraines are present along the valley walls, showing the former ice elevation and volume within the valley (Figure 6). These moraines are sloping downward toward the head of the valley showing greater ice thickness at the mouth of Moraine Canyon than near the head of the valley. Due to the geometry of adjoining glaciers, glacial ice levels in Moraine Canyon have fluctuated in the past based on levels of Amundsen Glacier at the mouth of the valley. Between January-August 2011, data loggers recorded mean air temperature in Moraine Canyon at -25.2°C, with maximum and minimum air temperatures for that time being -3.3°C and -44.8°C respectively. The debris covered ice which comprises most of the valley floor is flowing at a rate of 0.7 m/yr. Winds are predominantly from the polar plateau of the EAIS down into the valley, with some accumulation resulting from snow blown down from the Nilsen Plateau (Mayewski, 1975).
Figure 4. Location of Moraine Canyon in the Transantarctic Mountains. Map modified from Polar Geospatial Center, University of Minnesota.
Figure 5. Moraine Canyon aerial image. Red squares indicate sample locations. Valley length is ~25 km. Image shows both the debris covered ice and debris free ice in the Moraine Canyon flowing into Amundsen Glacier (not pictured) towards the bottom of the image. The inset map shows the relative location of Moraine Canyon within the Transantarctic Mountains.
Amundsen Glacier acts as a dam that based on its ice levels, prevents or allows the outflow of ice from Moraine Canyon. An effect of the increase in ice volume during Amundsen highstands is that ice from Amundsen flows up into Moraine Canyon and prevents the outflow of glacial ice from the valley. When Amundsen ice levels are lower, Moraine Canyon is no longer blocked by ice at the mouth of the valley. This allows the ice in Moraine Canyon to flow out of the valley and into Amundsen Glacier. The effects of lower Amundsen ice levels are analogous to pulling the drain plug out of a sink or bathtub. If ice levels in Amundsen are low enough then all of the glacial ice in Moraine Canyon is able to drain out of the valley.

Figure 6. Lateral moraines in Moraine Canyon. Red arrows point to several moraine crests on the left valley wall.
Distinct flow patterns are seen on the debris covered ice in the form of lobate debris lines showing faster ice flow in the center of the valley. The debris covered ice is also dotted with frozen meltwater ponds of unknown origins. Well defined arcing debris patterns on the dirty ice surface are deformed and stretched where they merge with the clean ice side. They appear to show forward and reverse motion of the debris covered side, suggesting that the ice flow changes during periods of thickening and thinning.

The bedrock geology of the Nilsen Plateau area consists of Precambrian metamorphic basement rocks overlain by dolerites, sandstones, granites, limestones, shales, and tillites (Mayewski, 1975). Valley walls are composed of unconsolidated regolith, and well preserved moraines are clearly visible along the exposed valley walls. Previous studies in the area of Moraine Canyon identified three main glacial drifts with successively older ages and higher elevations: Low, Middle, and High moraines (Mayewski, 1975). The highest elevation moraine sampled is at 1457 m, approximately 127 m above the current ice surface. The prevalence of multiple tiers of moraines and the large amount of exposed regolith make Moraine Canyon an exceptional location for exposure age determination and regolith degradation analysis.

**Methods**

**Field Procedures**

Sample pits were hand dug in lateral moraines on the valley walls (MC-PIT-13,14,15) and on the valley floor (MC-PIT-16,17). Regolith samples were collected at roughly 10cm increments to a maximum depth within the pits (Figure 6). Topographic shielding was measured in order to accurately calculate local nuclide production rates at each sample site (Dunne et. al, 1999). A compass and clinometer were used to take topographic shielding measurements, and CRONUS Earth online calculator was used to calculate the topographic shielding factors. In the field, latitude/longitude and elevation were measured with handheld GPS units, and air pressure
was measured with a handheld digital barometer. All measurements were normalized to a base camp GPS and barometer reading.

![Soil pit from lateral moraine, MC-PIT-14](image)

**Figure 7.** Soil pit from lateral moraine, MC-PIT-14. Total pit depth here is 43 cm.

**Laboratory Procedures**

Bulk regolith samples were prepared for AMS (accelerator mass spectrometry) at the University of North Dakota cosmogenic isotope laboratory using standard methods (Kohl and Nishiizumi, 1992; Stone, 2004; Balco and Shuster, 2009a; Balco and Shuster, 2009b). Density of bulk regolith was measured in order to determine the nuclide production rate at depth. Regolith samples were sieved to obtain 250-500µm sized particles, and quartz was separated out using heavy liquid lithium heteropolytungstate (LST). Isolated quartz samples were then repeatedly etched in 2% HF to remove any meteoric nuclides present. Samples were sent to PRIME laboratory at Purdue University for AMS measurements of $^{27}$Al/$^{26}$Al and $^{10}$Be/$^{9}$Be ratios and Berkeley Geochronology Center (BGC) for $^{21}$Ne noble gas spectrometer analysis. Carrier spikes
of beryllium were used in all of the samples, and aluminum carrier was added to samples with low aluminum concentrations.

$^{27}\text{Al}/^{26}\text{Al}$ and $^{10}\text{Be}/^{9}\text{Be}$ isotope ratios measured at PRIME Lab were referenced to NIST 2000. $^{21}\text{Ne}$ isotope concentrations measured at BGC were compared to quartz standards CREU-1 and CRONUS-A, reference materials produced from vein-quartz in the Atacama Desert in Chile and Antarctic sandstone respectively (Vermeesch et al., 2012). The $^{10}\text{Be}$ half-life used in this paper is 1.36 Mya with a corresponding decay constant of $\lambda = 5.1 \times 10^{-7} \text{yr}^{-1}$, and $^{26}\text{Al}$ half-life of 0.708 Mya with $\lambda = 9.78 \times 10^{-7} \text{yr}^{-1}$ (Nishiizumi, 2004; Nishiizumi et al., 2007). Site specific surface production rates of the three nuclides were calculated using sample densities, topographic shielding, elevation, and a sea-level, high-latitude reference production rate for $^{10}\text{Be}$ of 4.16 atoms g$^{-1}$ quartz (personal communication Daniel Morgan, 2016) after Stone (2000) and Balco and Shuster (2009b). All nuclide production in the top meter of regolith is assumed to be through spallation only (Goose and Phillips, 2001). Production rates and shielding data are shown in Table 1. Reported concentrations of $^{26}\text{Al}$, $^{10}\text{Be}$, and $^{21}\text{Ne}$, effective shielding, and sample densities are found in Table 2.

Table 1. Site Specific Nuclide Production Rates

<table>
<thead>
<tr>
<th>Sample Site</th>
<th>Latitude (°S)</th>
<th>Longitude (°E)</th>
<th>Pressure (hpa)</th>
<th>Shielding Correction</th>
<th>$P_{10}$ (at g$^{-1}$ yr$^{-1}$)</th>
<th>$P_{26}$ (at g$^{-1}$ yr$^{-1}$)</th>
<th>$P_{21}$ (at g$^{-1}$ yr$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MC-PIT-13</td>
<td>86.09606</td>
<td>157.56917</td>
<td>822</td>
<td>0.944</td>
<td>18.45</td>
<td>124.56</td>
<td>75.29</td>
</tr>
<tr>
<td>MC-PIT-14</td>
<td>86.10665</td>
<td>157.94720</td>
<td>827</td>
<td>0.960</td>
<td>17.87</td>
<td>120.62</td>
<td>72.91</td>
</tr>
<tr>
<td>MC-PIT-15</td>
<td>86.10696</td>
<td>157.96628</td>
<td>816</td>
<td>0.988</td>
<td>19.24</td>
<td>129.9</td>
<td>78.52</td>
</tr>
<tr>
<td>MC-PIT-16</td>
<td>86.09747</td>
<td>157.78178</td>
<td>831</td>
<td>0.997</td>
<td>17.41</td>
<td>117.52</td>
<td>71.03</td>
</tr>
<tr>
<td>MC-PIT-17</td>
<td>86.09750</td>
<td>157.78217</td>
<td>830</td>
<td>0.997</td>
<td>17.44</td>
<td>117.52</td>
<td>71.14</td>
</tr>
</tbody>
</table>
Table 2. Depth Density, Effective Shielding Mass, and Isotope Data. Dashed lines indicate no measurement for that individual sample.
Reported $^{21}\text{Ne}$ concentrations are concentrations from aliquot (a) from multiple aliquot analysis. Concentrations for all $^{21}\text{Ne}$ aliquots are found in table 5.

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Depth in Soil (cm)</th>
<th>Density (g/cm$^3$)</th>
<th>Effective Shielding Mass (g/cm$^2$)</th>
<th>$^{10}\text{Be} \times 10^6 \pm 1$ std (at g$^{-1}$ quartz)</th>
<th>$^{26}\text{Al} \times 10^6 \pm 1$ std (at g$^{-1}$ quartz)</th>
<th>$^{21}\text{Ne} \times 10^6 \pm 1$ std (at g$^{-1}$ quartz)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MC-PIT-13: right lateral moraine</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MC-PIT-13-4-8</td>
<td>4-8</td>
<td>1.96</td>
<td>11.79</td>
<td>9.091 ± 0.7236</td>
<td>-------</td>
<td>144.57 ± 2.9</td>
</tr>
<tr>
<td>MC-PIT-13-17-20</td>
<td>17-20</td>
<td>1.99</td>
<td>36.43</td>
<td>11.96 ± 0.8772</td>
<td>-------</td>
<td>153.51 ± 3.3</td>
</tr>
<tr>
<td>MC-PIT-14: left lateral moraine</td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>MC-PIT-14-3-7</td>
<td>3-7</td>
<td>1.66</td>
<td>8.78</td>
<td>7.587 ± 0.1967</td>
<td>41.57 ± 4.212</td>
<td>78.30 ± 1.8</td>
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<tr>
<td>MC-PIT-14-16-21</td>
<td>16-20</td>
<td>1.82</td>
<td>32.53</td>
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<td>-------</td>
<td>102.9 ± 2.2</td>
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<tr>
<td>MC-PIT-14-39-43</td>
<td>39-43</td>
<td>1.80</td>
<td>72.14</td>
<td>7.252 ± 0.2102</td>
<td>37.06 ± 2.765</td>
<td>89.0 ± 2.3</td>
</tr>
<tr>
<td>MC-PIT-15: left lateral moraine</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MC-PIT-15-2-6</td>
<td>2-6</td>
<td>1.73</td>
<td>7.38</td>
<td>16.15 ± 0.2760</td>
<td>77.72 ± 5.883</td>
<td>149.2 ± 3.7</td>
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<tr>
<td>MC-PIT-15-19-28</td>
<td>19-28</td>
<td>1.89</td>
<td>43.38</td>
<td>10.20 ± 0.1932</td>
<td>52.68 ± 15.29</td>
<td>90.5 ± 2.2</td>
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<tr>
<td>MC-PIT-15-40-46</td>
<td>40-46</td>
<td>1.92</td>
<td>79.5</td>
<td>9.906 ± 0.1881</td>
<td>62.77 ± 6.100</td>
<td>102.5 ± 2.7</td>
</tr>
<tr>
<td>MC-PIT-16: valley floor</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MC-PIT-16-1-7</td>
<td>1-7</td>
<td>1.87</td>
<td>7.51</td>
<td>-------</td>
<td>-------</td>
<td>80.8 ± 1.9</td>
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<tr>
<td>MC-PIT-16-10-14</td>
<td>10-14</td>
<td>1.84</td>
<td>23.59</td>
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<td>85.5 ± 2.5</td>
</tr>
<tr>
<td>MC-PIT-16-20-24</td>
<td>20-24</td>
<td>1.99</td>
<td>43.45</td>
<td>-------</td>
<td>-------</td>
<td>79.3 ± 2.2</td>
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<tr>
<td>MC-PIT-16-29-30</td>
<td>29-30</td>
<td>1.86</td>
<td>57.64</td>
<td>-------</td>
<td>-------</td>
<td>81.1 ± 1.9</td>
</tr>
<tr>
<td>MC-PIT-17: valley floor</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MC-PIT-17-0-5</td>
<td>0-5</td>
<td>1.86</td>
<td>4.67</td>
<td>9.682 ± 0.2317</td>
<td>54.96 ± 3.679</td>
<td>86.14 ± 2.4</td>
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<tr>
<td>MC-PIT-17-13-18</td>
<td>13-18</td>
<td>1.96</td>
<td>29.11</td>
<td>9.254 ± 0.3142</td>
<td>58.12 ± 33.44</td>
<td>88.26 ± 2.1</td>
</tr>
<tr>
<td>MC-PIT-17-28-32</td>
<td>28-32</td>
<td>1.88</td>
<td>56.38</td>
<td>-------</td>
<td>-------</td>
<td>85.58 ± 2.2</td>
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<tr>
<td>MC-PIT-17-42-45</td>
<td>42-45</td>
<td>1.83</td>
<td>81.77</td>
<td>8.496 ± 0.1809</td>
<td>50.26 ± 5.441</td>
<td>79.41 ± 2.1</td>
</tr>
</tbody>
</table>
**Exposure Model**

Calculating the exposure ages and erosion rates of regolith deposits is complicated by the possibility that the regolith was deposited with an inherited nuclide concentration from prior periods of exposure. Inherited nuclide concentrations will produce a steeper depth-concentration profile than the expected post depositional profile. By measuring nuclide concentrations along a vertical depth profile, nuclide inheritance can be constrained and a more accurate exposure age and erosion rate can be determined.

With a steady, non-eroding surface, the concentration of a given nuclide, $N_i$, (atoms g$^{-1}$ quartz) under a given shielding mass, $z$, (g cm$^{-2}$) and exposure time, $t$, (years) can be described by Equation 1 (Lal, 1991).

$$N_i(z,t) = N_{inh}(z)e^{-t \lambda} + \frac{P_i(z)}{\lambda}(1 - e^{-t \lambda})$$

The subscript $i$ refers to the nuclide of interest, $N_{inh}$ is the inherited nuclide concentration that remains in the sample today (atoms g$^{-1}$ quartz), $P_i(z)$ is the nuclide production rate at a given depth (atoms g$^{-1}$ quartz yr$^{-1}$). This equation can be simplified for stable $^{21}$Ne by removing the terms with decay constants via Equation 2 (Niedermann, 2002).

$$N_i(z,t) = N_{inh}(z) + P_i(z)t$$

For radionuclides $^{26}$Al and $^{10}$Be, secular equilibrium is reached when the nuclide production rate equals the decay rate. At this point, only degradation or aggradation will affect the concentration of the radionuclides. Once secular equilibrium is reached, the nuclide concentrations provide only absolute minimum exposure ages. The time required to reach this point is known as the effective half-life. This can be determined after via Equation 3 where $\tau_{1/2,e}$ is the effective half-life, and $\varepsilon$ is the regolith degradation rate (g cm$^{-2}$ yr$^{-1}$).
The attenuation length ($\Lambda$) is taken to be 150 g cm$^{-2}$ for Antarctica. See Gosse and Phillips (Gosse and Phillips, 2001) for detailed discussion of $\Lambda$. Degradation rates are calculated after a few effective half-lives have passed.

The expected concentration of a nuclide ($N_{i,n}$) from a sample experiencing steady erosion is described by Equation 4 (Lal, 1991; Niedermann, 2002).

$$N_{i}(\varepsilon, t) = N_{i,inb} + \frac{P_i \cdot e^{-x_j/\Lambda}}{\lambda_i + \varepsilon/\Lambda} \cdot (1 - e^{-(\lambda_i + \varepsilon/\Lambda) \cdot t})$$

The subscript $j$ is the individual sample from a site, $\varepsilon$ is the degradation rate (g cm$^{-2}$ yr$^{-1}$), and $P_i$ is the production rate of the nuclide of interest at the surface (atom g$^{-1}$ quartz yr$^{-1}$). Degradation rates are calculated as the amount of shielding mass removed per year then converted to more common units of m Ma$^{-1}$.

The unknown parameters in these equations are $t$, $\varepsilon$, and $N_{i,n}$. Using an optimization model in MATLAB® (Morgan et al., 2010b), we find the combination of these unknown parameter values that produce the least difference between the observed and modeled nuclide concentration profiles. The observed and modeled concentration profiles were compared using standard error weighted chi squared ($\chi^2$) minimization techniques (Braucher et al., 2009; Morgan et al., 2010b).

The best modeled fit is determined by minimizing the $\chi^2$ value; therefore the smaller the $\chi^2$ value the better the fit. Modeled best fit nuclide concentrations are calculated from initial parameter estimations, and lower and upper boundary conditions are set for time, erosion rate, and inherited nuclide concentrations to eliminate local minima and maxima. Boundary conditions for these parameters are as follows: inherited nuclides min=0, max=nuclide saturation.
(no max or min are set for stable $^{21}$Ne); erosion min=0, max =infinite; time min=0, max=5 Mya, which is duration of a few effective half-lives.

The exposure ages of the deposits can be calculated from the modeled surface concentrations fit to the nuclide depth profile, or they can be calculated using a single nuclide concentration at depth. In the latter method, $T_{exp}$ represents the apparent exposure age following Equation 5 (Lal, 1991; Niedermann, 2002).

$$T_{exp} = -\left(\frac{1}{\lambda}\right) \cdot \ln(1 - \frac{N(z) \cdot \lambda}{P(z) \cdot e^{-\frac{\rho \cdot z}{\lambda}}})$$

The total nuclide concentration in a sample, $N$, can be further modified to account for inherited nuclides if they are able to be determined independently from the nuclide depth profile. The effect of erosion on a sample can be taken into account using Equation 6 which determines an erosion scaling factor ($f_e$) that is then multiplied by $T_{exp}$ to calculate the erosion corrected exposure age.

$$f_e = 1 + \frac{e^{T_{exp} \cdot \rho}}{z}$$

**Error Analysis**

To determine uncertainties in the modeled results, we used a 10,000 run Monte Carlo simulation which takes into account the uncertainties in measured nuclide concentrations. This technique assumes a Gaussian distribution with one-sigma errors for the measured nuclide concentrations (Balco et al., 2005; Morgan et al., 2010b). For each run of the Monte Carlo simulation, a random concentration within the measurement error is selected and best fit parameters are fit by minimizing $\chi^2$. The simulation produces a histogram of possible solutions for the unknown variables in the exposure model. The error of the best fit parameters is a standard 1σ (68%) of the 10,000 run simulation.
Results

Local geologic setting and observations are paired with the analysis of cosmogenic nuclide concentrations to determine the occurrence and timing of geologic events in Moraine Canyon. Measured concentrations of $^{10}\text{Be}$ and $^{26}\text{Al}$ are modeled together to attain a best solution for exposure age, erosion rate, and nuclide inheritance, and $^{21}\text{Ne}$ concentrations are used to better constrain exposure ages.

High percentages of inherited nuclides are expected for all samples because the regolith is assumed to have had prior periods of exposure and has been reworked from previous deposits. The exposure model assumes that the regolith was well mixed when it was deposited *in-situ*, meaning the concentration of inherited nuclides is expected to be the same at all depths. $^{21}\text{Ne}$ is a stable isotope that doesn’t experience radioactive decay. Its concentration in a sample represents the current nuclide accumulation as well as all previous accumulations. The concentration of the deepest sample in each pit may be assumed to be the maximum allowed inheritance for $^{21}\text{Ne}$.

The simplest interpretation of the nuclide data is the determination of an exposure age assuming zero erosion and zero nuclide inheritance; however this is simply a starting point for the age analysis. Solutions for exposure age, erosion rate, and nuclide inheritance are calculated using modeled concentrations fit to the measured depth profile. The minimum modeled exposure age (E=0, Inh=best) is calculated in order to better constrain the actual age of each deposit. The best modeled exposure age, assuming the best modeled erosion rate and inheritance values is reported here. Also included in the results is the apparent exposure age of the deposits ($T_{exp}$) calculated from the top-most measured nuclide concentration and Equation 5. Figures 7-10 show measured nuclide concentrations within each pit plotted with the best fit concentration line.

Erosion rates, inherited nuclide concentrations and exposure ages are shown in Tables 3-4. The
results for each deposit are discussed in more detail in the following sections ordered from oldest to youngest deposit.

**PIT 15 Moraine**

PIT 15 is located on a prominent, flat topped lateral moraine on the right side of the valley. Concentrations of $^{10}$Be, $^{26}$Al, and $^{21}$Ne were measured for PIT 15. Nuclide concentrations in this sample are higher than all of the other samples, and the PIT-15 moraine is at the highest sampled elevation. The color here transitions from black rock of the deposit below to caramel browns on the terrace and above suggesting a higher degree of *in-situ* weathering. This difference in coloration indicates that this deposit is potentially older than the deposit below. The nuclide concentrations, elevation, and degree of weathering support the expectation that the PIT-15 moraine is older than the other deposits. Depth profiles of all three nuclides show overall decreasing concentration with depth (Figure 8).

The calculated erosion rate is $0.21 \pm 0.20/\pm 0.06$ m Ma$^{-1}$ and the inherited nuclide concentrations remaining in the sample today are $^{10}$Be $N_{\text{inh}} = 5.58 \pm 5.50/\pm 6.42 \times 10^5$ atoms g$^{-1}$ quartz and $^{26}$Al $N_{\text{inh}} = 3.74 \pm 3.69/\pm 4.14 \times 10^6$ atoms g$^{-1}$ quartz. The minimum modeled exposure age of this moraine is $1.04 \times 10^6$ years, and the best modeled exposure age is $1.21 \pm 2.07/\pm 1.08 \times 10^5$ years. When corrected for best modeled erosion and inheritance values, the apparent exposure age is $T_{\text{exp,E}=0.21,\text{Inh(Al)}} = 1.13 \times 10^6$, $T_{\text{exp,E}=0.21\text{Inh(Be)}} = 1.40 \times 10^6$, and $T_{\text{exp,E}=0.21,\text{Inh(Ne)}} = 7.74 \times 10^5$ years.
PIT 13 Moraine

PIT 13 is located on the steep sided moraine on the right side of the valley. The moraine is composed mostly of boulders and depth samples were difficult to obtain. Two subsurface samples were collected for this pit and concentrations of $^{10}\text{Be}$ and $^{21}\text{Ne}$ were measured (Figure 9). The difficulty of collecting samples may have led to the mixing and contamination of separate samples within the pit; therefore the samples from PIT 13 are of questionable quality. Because the concentrations do not follow the expected concentration-depth relationship, this exposure model cannot be used to determine erosion rates, inherited nuclide concentrations, or ages between individual depth samples. Instead, age and erosion rate solutions for PIT 13 were calculated using Equation 5 and the top-most measured concentrations of $^{10}\text{Be}$ and $^{21}\text{Ne}$. Inherited nuclide concentrations were not able to be determined for this location; therefore the resulting apparent exposure age is a maximum possible age rather than an actual exposure age. Only one radionuclide was used in the analysis of PIT 13 therefore a steady state uniform erosion rate is assumed for the surface. The calculated steady state erosion rate is 1.12 m Ma$^{-1}$. Assuming steady state erosion, the corrected apparent exposure age of the PIT-13 moraine is $T_{\text{exp,E}=1.12(\text{Be})} = 1.10 \times 10^6$ years and $T_{\text{exp,E}=1.12(\text{Ne})} = 5.91 \times 10^6$ years.

PIT 16, 17 Deposit

PIT 16 and 17 are located on the valley floor which is composed of regolith-covered glacial ice. The lithology is varied suggesting origins of the regolith are widespread throughout the valley or possibly outside the valley. Concentrations of $^{21}\text{Ne}$ were measured for PIT 16, and all three nuclides were measured for PIT 17 (Figure 10). Underlying ice was found at depths of
29 cm and 42 cm for PIT 16 and 17 respectively. Ice was visibly clean with a few small dirt filled cracks.

The $^{10}$Be and $^{26}$Al concentration profiles are consistent with stable, non-mixed regolith profiles. The calculated erosion rate is $4.40\pm0.23$ m Ma$^{-1}$, and the inherited nuclide concentrations remaining in the sample today are $^{10}$Be $N_{inh}=8.13\pm1.23 \times 10^5$ atoms g$^{-1}$ quartz and $^{26}$Al $N_{inh}=4.87\times10^7\pm1.12 \times 10^5$ atoms g$^{-1}$ quartz. The minimum modeled exposure age of this moraine is $1.78 \times 10^5$ years, and the best modeled exposure age is $2.87\pm1.16 \times 10^5$ years. When corrected for best modeled erosion and inheritance values, the apparent exposure age of PIT-17 is $T_{exp17, E=4.40, Inh(Al)}=1.46 \times 10^5$, $T_{exp17, E=4.40, Inh(Be)}=2.49 \times 10^5$, and $T_{exp17, E=4.40, Inh(Ne)}=3.68 \times 10^5$ years. Apparent exposure age from PIT-16 $^{21}$Ne concentrations corrected for modeled erosion rate from PIT-17 is $T_{exp16, E=4.40, Inh(Ne)}=1.68 \times 10^5$ years.

**PIT 14 Moraine**

PIT 14 was collected on a lower lateral moraine on the left side of the valley. This deposit is dominated by 3-5 cm clasts and surrounded by larger 0.5-1 m boulders. Concentrations of $^{10}$Be, $^{26}$Al, and $^{21}$Ne were measured for PIT 14 (Figure 11). $^{10}$Be and $^{26}$Al concentrations are the lowest in this pit compared to the others suggesting a shorter exposure time. $^{21}$Ne concentrations in PIT 14 do not show any systematic relationship with depth, therefore it is assumed that the majority of the $^{21}$Ne concentration is inherited from prior exposure. The regolith has not been in-situ for long enough to for the in-situ production of $^{21}$Ne to overcome the inherited $^{21}$Ne concentration. One possibility is that some vertical mixing of the regolith has occurred, however the $^{10}$Be and $^{26}$Al profiles exhibit the expected nuclide concentration-depth relationship. The
steep concentration vs. depth profile suggests that there is a high concentration of inherited nuclides from prior exposure (Figure 11).

The calculated erosion rate is 4.16 -4.30/+0.23 m Ma\(^{-1}\), and the inherited nuclide concentrations remaining in the sample today are \(^{10}\text{Be}\) \(N_{\text{inh}}= 6.78 \times 10^6 -4.27/+5.29 \times 10^5\) atoms g\(^{-1}\) quartz and \(^{26}\text{Al}\) \(N_{\text{inh}}= 3.56 \times 10^7 -2.92/+3.85 \times 10^6\) atoms g\(^{-1}\) quartz. The minimum modeled exposure age of this moraine is 6.61 \(\times 10^4\) years, and the best modeled exposure age is 8.01 -5.94/+6.94 \(\times 10^4\) years. When corrected for best modeled erosion and inheritance values, the apparent exposure age is \(T_{\text{exp, E=4.16, Inh(Al)}} = 8.48 \times 10^4\), \(T_{\text{exp, E=4.16, Inh(Be)}} = 7.88 \times 10^4\), and \(T_{\text{exp, E=4.16, Inh(Ne)}} = 4.06 \times 10^5\) years.
Figure 8. Nuclide Concentrations vs. Effective Shielding Mass for PIT-15. PIT-15 is located on a lateral moraine on the left valley wall looking down valley. Measured concentrations of $^{10}$Be (circles), $^{26}$Al (squares), and $^{26}$Ne (triangles) include vertical lines showing depth range and horizontal lines showing measurement error for each sample. Some measurement errors are too small to be shown on this scale. Solid lines show best fit for each nuclide from erosion model, and dashed lines show best fit inherited nuclide concentrations. $^{21}$Ne concentrations include all measured concentrations from multiple aliquots of each sample.
Figure 9. Nuclide Concentrations vs. Effective Shielding Mass for PIT-13. PIT-13 is located on the right valley wall looking down valley. Measured concentrations of $^{10}\text{Be}$ (circles) and $^{26}\text{Ne}$ (triangles) include vertical lines showing depth range and horizontal lines showing measurement error for each sample. Measurement error for $^{21}\text{Ne}$ is too small to be shown on this scale. No best fit line was determined because concentration-depth profile does not follow the decreasing concentration with decreasing depth trend. Inheritance values were not able to be determined for PIT-13. $^{21}\text{Ne}$ concentrations include all measured concentrations from multiple aliquots of each sample.
Figure 10. Nuclide Concentrations vs. Effective Shielding Mass for PIT-16 and PIT-17. These pits are both located in close proximity on the valley floor. Measured concentrations of $^{10}\text{Be}$ (circles), $^{26}\text{Al}$ (squares), and $^{26}\text{Ne}$ (triangles) include vertical lines showing depth range and horizontal lines showing measurement error for each sample. Some measurement errors are too small to be shown on this scale. Solid lines show best fit for each nuclide from erosion model, and dashed lines show best fit inherited nuclide concentrations. $^{21}\text{Ne}$ concentrations include all measured concentrations from multiple aliquots of each sample.
Figure 11. Nuclide Concentrations vs. Effective Shielding Mass for PIT-14. PIT-14 is located on a lateral moraine on left valley wall looking down valley. Measured concentrations of $^{10}\text{Be}$ (circles), $^{26}\text{Al}$ (squares), and $^{26}\text{Ne}$ (triangles) include vertical lines showing depth range and horizontal lines showing measurement error for each sample. Some measurement errors are too small to be shown on this scale. Solid lines show best fit for each nuclide from erosion model, and dashed lines show best fit inherited nuclide concentrations. $^{21}\text{Ne}$ concentrations include all measured concentrations from multiple aliquots of each sample.
Table 3. Modeled Erosion Rate, Nuclide Inheritance, Exposure Age, and $\chi^2$. Dashed lines indicate no measurement for that individual sample.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Erosion Rate (m Ma$^{-1}$)</th>
<th>$^{26}$Al inheritance (at g$^{-1}$ quartz)</th>
<th>$^{10}$Be inheritance (at g$^{-1}$ quartz)</th>
<th>$^{21}$Ne inheritance (at g$^{-1}$ quartz)</th>
<th>Minimum Exposure Age (E=0, Inh=best)</th>
<th>Best Exposure Age (E=best, Inh=best)</th>
<th>$\chi^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>MC-PIT-13</td>
<td>1.12</td>
<td>-----</td>
<td>------</td>
<td>------</td>
<td>6.83 x 10$^5$</td>
<td>1.10 x 10$^6$</td>
<td>------</td>
</tr>
<tr>
<td>MC-PIT-14</td>
<td>4.16</td>
<td>3.57 x 10$^7$</td>
<td>6.78 x 10$^6$</td>
<td>6.51 x 10$^7$</td>
<td>6.61 x 10$^4$</td>
<td>8.04 x 10$^4$</td>
<td>0.18</td>
</tr>
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<td>0.211</td>
<td>3.74 x 10$^6$</td>
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<td>1.02 x 10$^8$</td>
<td>1.04 x 10$^6$</td>
<td>1.21 x 10$^6$</td>
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<td>-----</td>
<td>------</td>
<td>7.65 x 10$^7$</td>
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<td>MC-PIT-17</td>
<td>4.4</td>
<td>4.87 x 10$^7$</td>
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<td>0.85</td>
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Table 4. Apparent Exposure Ages. Apparent exposure ages are calculated using Equation 5, the measured nuclide concentration in the top-most pit sample, and the modeled erosion rate and modeled nuclide inheritance. Dashed lines indicate no measurement for that individual sample.

<table>
<thead>
<tr>
<th>Sample</th>
<th>$^{26}$Al Apparent Exposure Age (years)</th>
<th>$^{10}$Be Apparent Exposure Age (years)</th>
<th>$^{21}$Ne Apparent Exposure Age (years)</th>
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<td>MC-PIT-16</td>
<td>-----</td>
<td>-----</td>
<td>1.68 x 10$^5$</td>
</tr>
<tr>
<td>MC-PIT-17</td>
<td>1.46 x 10$^5$</td>
<td>2.49 x 10$^5$</td>
<td>3.68 x 10$^5$</td>
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</tbody>
</table>
Figure 12. Sample locations and best exposure ages. Exposure ages with best modeled erosion rate and inherited nuclide concentrations applied. * denoted apparent exposure age from only $^{21}$Ne concentrations. ** denotes absolute maximum exposure age with $E=\text{best}$, $\text{Inh}=0$. 
Discussion

Moraine Canyon is a dynamic location that has experienced numerous fluctuations in ice level that are controlled by fluctuations in thickness of Amundsen Glacier. Distinct lateral moraines in Moraine Canyon become increasingly older with increasing elevation, suggesting that earlier expansions of glacial ice in the valley were more extensive than later expansions. If they occurred, larger subsequent ice expansions would have overrun any lower glacial deposits. The decreasing exposure age with decreasing elevation and the currently low ice levels in Moraine Canyon provide evidence for an overall decrease in ice volume and thickness of Amundsen over the past >1.21 Ma. This decrease in ice levels is consistent with evidence for the gradual thinning of the EAIS over the past few million years and is seen in other locations throughout the continent (e.g. Fogwill et al., 2004; Altmaier et al., 2010; Lilly et al., 2010; Swanger et al., 2011). The overall decline in ice was punctuated by several shorter term periods of expansion and thickening as evidenced by the lateral moraines below the PIT-15 moraine on the valley walls.

The orientation and slanting of the lateral moraines show that they were deposited during periods of higher ice levels in Amundsen Glacier. The moraines are sloping down toward the head of the valley, showing there was greater ice thickness at the mouth of valley than at the head of the valley due to the backing up of ice from Amundsen Glacier. This pattern of asymmetric ice thickening has been observed at EAIS outlet glaciers throughout the TAM where thickening of the WAIS causes ice thickening to propagate upstream (Bockheim et al., 1989; Bromley et al., 2010; Todd et al., 2010).

The exposure ages of the moraines indicate periods of ice thickening in Amundsen which is a result of the expansion of the EAIS, WAIS, or both. Ice highstands are recorded by the lateral moraines on the valley walls, but periods of ice thinning leave no in-situ geologic evidence for interpretation. In Moraine Canyon, the age relationship between the youngest
sampled lateral moraine (80.4 ka) and the supraglacial deposit on the valley floor (287 ka) provides evidence for at least two periods of ice expansion following a period of significant thinning and almost complete clearing of glacial ice from the valley. The sequence of ice level changes is described with the following chronology from oldest to youngest deposit.

1) The PIT-15 moraine (1.21 Ma) was deposited during a highstand of Amundsen Glacier which blocked the ice in Moraine Canyon. Ice from Amundsen flowed back into Moraine Canyon as a result of the increased ice volume in Amundsen. After this ice thickening event, Amundsen ice levels decreased, allowing ice levels in Moraine Canyon to decrease as well.

2) Another thickening event occurred in Amundsen Glacier which caused the backing up of ice from Amundsen into Moraine Canyon again and resulted in the deposition of the PIT-13 moraine at (1.10 Ma). This moraine is lower in elevation which suggests that the ice expansion in the valley was not as extensive as the previously dated ice expansion.

3) After the thickening event during which the PIT-13 moraine was deposited, there was a period of major ice lowering of Amundsen Glacier. This significant drop in ice levels at the mouth of the Moraine Canyon allowed the majority of ice in the valley to flow out and into Amundsen. There are no deposits present from this ice lowstand because subsequent ice expansions have overrun any lower deposits. This thinning event is discussed in detail in the following sections.

4) With Moraine Canyon free or almost free of ice after this period of intense thinning of Amundsen, the next expansion of Amundsen Glacier pushed ice and fresh regolith back into the valley. Unconsolidated regolith was deposited on the surface of the ice
eventually producing the PIT-16, 17 deposit on the valley floor (287 Ka). There are no dated lateral moraines of this age on the valley walls, and that may be due to the fact that not all of the moraines in the valley were sampled or that later ice expansions removed any deposits from this time. In order for glacial ice to be reestablished in Moraine Canyon after it was ice-free, a large volume of ice would have to flow into the valley from Amundsen during the next ice thickening event. Any lateral moraines deposited during the thickening event at 287 ka would then have been overrun by subsequent ice thickening events.

5) The PIT-14 moraine was deposited during another Amundsen Glacier thickening event 80 ka. This moraine is the lowest in elevation which suggests that there was less ice thickening during this time than during the PIT 15 and PIT 13 expansions, but most likely more ice thickening than during the PIT 16, 17 expansion that reestablished glacial ice in Moraine Canyon.

6) Currently, ice levels in Moraine Canyon are low and ice is flowing down valley to Amundsen. Unique arcing debris patterns on the debris covered ice show this back and forth movement of the underlying ice since the supraglacial debris was deposited. However, the forward and reverse movement of the ice has not been significant enough to disturb or considerably mix the overlying regolith in the upper parts of the valley.

**Interpretation of valley floor deposit**

My interpretation of an ice draining event occurring prior to 287 ka is supported by the nuclide concentration profiles in PIT 16 and PIT 17 from the valley floor. The stable profiles show no vertical mixing of the regolith as the ice movement changes direction, therefore the
regolith is being transported back and forth within the valley atop the glacial ice. The high concentrations of inherited nuclides in these pits support the idea that the regolith was reworked from another deposit that had a prior period of exposure. Had there been no clearing of ice and regolith from the valley, the exposure age of the PIT 16, 17 deposit should be significantly older than 287 ka, and the age of the valley floor deposit should correlate to one of the lateral moraines on the valley wall. If this were the case, there would be little or no remaining concentration of inherited \(^{26}\text{Al}\) and \(^{10}\text{Be}\) because any nuclide signature from prior exposure would have decayed away. Instead we see high concentrations of inherited nuclides in PITS 16 and 17.

In order to produce the supraglacial deposit that dates to 287 ka and not significantly older, there needed to be a clearing event where older ice and regolith was completely drained from the valley and fresh ice and regolith was redeposited during a subsequent thickening event. This clearing event had to occur between the ice thickening events that deposited the lateral moraine at 1.10 Ma and the valley floor deposit at 287 ka. During this time, ice levels in Amundsen Glacier were low enough to allow all of the ice in Moraine Canyon to flow out of the valley, removing all of the old regolith leaving the valley ice free.

Another interpretation of the concentration profiles and exposure age is that 287 ka is the time when the regolith became stable and stopped mixing. As the glacial ice in Moraine Canyon goes through cycles of forward flow and backing up as a result of the fluctuations of Amundsen, the supraglacial debris may be disrupted and become well mixed. Therefore this exposure age could simply mark when the supraglacial debris stopped being repeatedly mixed as a result of the flow changes in Moraine Canyon. However, based on the stable concentration profiles from this deposit we see that there has been no mixing of the regolith since its deposition at 287 ka. While there has been no mixing of the regolith, there have been periods of ice thickening and thinning
that would have allowed ice to flow out of the valley as well as get pushed back into the valley since the deposition of the PIT 16, 17 deposit. This is evident by the younger PIT 14 lateral moraine which shows an ice thickening event at 80 ka as well as the downstream flow of ice occurring in Moraine Canyon today. If we consider similar conditions of repeated changes in ice flow and lack of regolith mixing prior to 287 ka, it is unlikely that this age would mark the time when the supraglacial debris became stable and stopped mixing because regolith mixing would not have been occurring. Therefore I have dismissed the possibility that 287 ka is the time when the supraglacial deposit became stable and stopped mixing.

**EAIS Thinning**

Because of Moraine Canyon’s location between the EAIS and WAIS, it is affected by fluctuations of both ice sheets. The current ice elevation at the confluence of Moraine Canyon and Amundsen glacier is ~1200 m., and this elevation is approximately the median of the current ice elevations between the WAIS (~200m) and the EAIS (~2200m) in the head of Amundsen Glacier. The overall decrease in ice levels in the valley are a result of the gradual thinning of the EAIS over the past few million years. Long term cooling of the climate since the warmer Pliocene results in less moisture and precipitation being transported to the polar plateau causing overall thinning of the EAIS (Altmaier et al., 2010; Suganuma et al., 2014). Periodic ice advances shown by the deposited lateral moraines in the valley are associated with WAIS fluctuations during shorter term glacial-interglacial cycles.

Considering the currently low WAIS level, we see that there is still a considerable amount of glacial ice present in Moraine Canyon; this is due to Amundsen Glacier being sufficiently thick enough to partially dam the flowing ice from Moraine Canyon. For the ice levels in Amundsen Glacier to drop enough to initiate the outflow of ice from Moraine Canyon, a
significant amount of ice thinning of the EAIS as well as the WAIS would be required. A
decrease in WAIS thickness alone would be insufficient to enable the draining of ice from the
valley. Simultaneous thinning of the EAIS and WAIS would enable ice levels in Amundsen
Glacier to be low enough to allow the majority of ice drain from Moraine Canyon. The shape and
dimensions of the valley suggests that while most of the ice was drained, a small body of
stagnant ice most likely remained on the valley floor.

Previous investigations suggest that the EAIS is reasonably stable and resilient to shorter
term climatic changes, but recent studies from off shore sediment cores (Theissen et al., 2003;
Reinardy et al., 2015), and this evidence from Moraine Canyon suggest that the EAIS is more
dynamic and sensitive to climatic changes over shorter periods of time. The ice thinning event
that allowed the ice to be cleared from Moraine Canyon occurred at some point after the
deposition of the PIT-13 moraine (1.10 Ma) and prior to the deposition of the PIT-16,17 deposit
on the valley floor (287 ka). This period of thinning may have occurred during a prolonged and
intense interglacial period, MIS 11 (420 ka), that falls within this timespan. Models of Antarctic
ice volume imply the full to partial collapse of the WAIS, ~420 ka (Pollard and DeConto, 2009),
and AND-1B sediment cores suggest a similar melting/collapse of the WAIS during MIS 7 (240
ka) (Naish et al., 2009; McKay et al., 2012). The geologic evidence from Moraine Canyon and
the connection to an intense interglacial period and possible collapse of the WAIS support the
interpretation that a period of significant WAIS and EAIS thinning occurred in the Amundsen
Glacier region 1.10 Ma-287 ka years. This interpretation supports a more dynamic EAIS than
previously thought.
Conclusions

In this study, I reconstruct the glacial history of Moraine Canyon, a tributary of Amundsen Glacier, over the past 1.21 Ma based on cosmogenic exposure dating of glacial deposits. Dated deposits show a decrease in exposure age with decreasing elevation. Exposure ages from glacial deposits in Moraine Canyon show an overall decrease in ice levels over the past >1.21 Ma. The overall decrease in ice levels is punctuated by at least three thickening events and an intense period of thinning during which ice was completely or nearly completely cleared out of Moraine Canyon. The key implication of this research is that a period of significant EAIS thinning occurred in the Amundsen Glacier region between ~1.21 Ma and 287 ka. During this thinning event, EAIS elevation decreased significantly, resulting in the significant lowering of Amundsen Glacier which allowed the ice in Moraine Canyon to almost completely flow out of the valley. In-situ geologic evidence present from lowstands of the ice sheets is rare because subsequent ice advances overrun any evidence of lower ice levels, but the regolith deposits on the valley walls and floor of Moraine Canyon provide evidence for a period of significantly lower EAIS and WAIS. The geologic and geomorphic evidence discussed here provides new constraints on previously undetermined EAIS fluctuations in the southern TAM.
CHAPTER III
CONCLUSIONS

The objective of this research using cosmogenic nuclide concentrations was to assign minimum ages to the glacial deposits and quantify regolith erosion rates in Moraine Canyon in the southern TAM. $^{26}\text{Al}$, $^{10}\text{Be}$, and $^{21}\text{Ne}$ nuclide concentrations were measured at different depths along a vertical profile within unconsolidated glacial deposits on the valley walls and valley floor. Reported exposure ages of the deposits are corrected for local erosion rates and inherited nuclide concentrations.

Continuous degradation of surficial deposits adds a level of complexity to paleoclimate interpretation, and nuclide concentrations provide more information than solely the exposure age of deposits. By analyzing nuclide concentrations along a vertical depth profile in regolith deposits we can better understand regolith transport, erosion rates, and nuclide inheritance within the deposits. Regolith erosion rates calculated here range from 0.21-4.40 m Ma$^{-1}$. These rates fall within the erosion rate range to almost double the erosion rates observed in the MDV at 0.19-2.6 m Ma$^{-1}$ (Putkonen et al., 2008a; Morgan et al., 2010b).

Overall nuclide concentrations of $^{26}\text{Al}$, $^{10}\text{Be}$, and $^{21}\text{Ne}$ in all pit locations are similar, suggesting that the system is dynamic and the regolith was well mixed at the time of deposition after being reworked several times. This is also evident by the high concentrations of inherited nuclides in all of the pits. The steep nuclide concentration vs. depth profile of $^{26}\text{Al}$, $^{10}\text{Be}$, and $^{21}\text{Ne}$ seen in Pits 14, 16, and 17 suggests that there is a high concentration of inherited nuclides from prior exposure. The Pit-15 moraine has been "in-situ" for long enough for
most of the inherited $^{26}\text{Al}$ and $^{10}\text{Be}$ to decay, showing a less steep depth profile. There is a significant time gap between the deposition of the Pit-15 moraine and the Pit-14 moraine. This paired with their proximity in elevation to one another suggests that there may have been periods of ice thickening between 1.21 Ma and 80 ka, but evidence for these periods has been overrun by subsequent ice expansions.

Concentrations of $^{10}\text{Be}$ and $^{21}\text{Ne}$ in PIT 13 do not follow the expected depth-concentration profile. Analytical error has been ruled out in this case because both nuclides follow the same trend of increasing concentration with depth. This would suggest some sort of vertical mixing has occurred or more likely mixing or contamination of sediment when the samples were collected.

Located adjacent to Amundsen Glacier, Moraine Canyon is a unique location that is affected by changes in both the EAIS and WAIS. The EAIS response to warming climate is more complicated that the WAIS response. The EAIS experiences increased ice mass loss through the thinning of outlet glaciers and increased surface accumulation in the interior of the ice sheet. The marine-based WAIS has responded more readily to glacial-interglacial fluctuations, and this is taken into account in ice sheet models (Pollard and DeConto 2009). Overall thinning of the EAIS since the early Pleistocene is caused by the reduction of moisture transport and precipitation from the Southern Ocean (Suganuma et al., 2014). This is seen in the progressively lower elevation and younger glacial drifts in Moraine Canyon. However, it is unclear if the EAIS behaves as dynamically as the WAIS.

Geologic evidence based on moraine limits on the inland side of the Transantarctic Mountains suggests little change in the surface of the East Antarctic Ice Sheet at the LGM (Bockheim et al., 1989; Denton et al., 1989a,b). This may suggest that the EAIS does not
respond to shorter duration changes in climate. However, recent evidence from oceanic sediment cores show a more dynamic EAIS near the continental margins (Theissen et al., 2003; Reinardy et al., 2015). One of the major findings of this work is that there was a period of significant EAIS thinning in the Amundsen Glacier region sometime between 1.21 Ma and 2.87 ka. These results suggest a more dynamic EAIS than previously thought.

With the current rise in greenhouse gases and global temperatures, studying polar climates and ice sheet dynamics is imperative to predicting future ice sheet behavior and sea-level changes. This research using cosmogenic nuclides provides evidence for past glacial changes, insight into ice dynamics, and information on regolith transport in the region where very few measurements currently exist. Understanding of glacial fluctuations and sediment transport will help to interpret climatic history and geomorphic processes in the southern TAM. The numerous glacial deposits in the Moraine Canyon make this an ideal location to study the glacial history of the EAIS. Due to monetary and time constraints in this study, samples from only a few deposits were taken and dated. A more comprehensive study of this or similar areas would provide further insights into WAIS and EAIS fluctuations.
APPENDIX

Table 5. $^{21}$Ne Concentrations. Reported concentrations include multiple aliquots for each sample.

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Depth in Soil (cm)</th>
<th>Effective Shielding Mass (g/cm$^2$)</th>
<th>Aliquot</th>
<th>$^{21}$Ne x 10$^6$ (at g$^{-1}$ quartz)</th>
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<td>MC-PIT-13: right lateral moraine</td>
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<td>MC-PIT-13-4-8</td>
<td>4-8</td>
<td>11.79</td>
<td>a</td>
<td>144.57 ± 2.9</td>
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<td>b</td>
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<td>153.51 ± 3.3</td>
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<td>154.02 ± 2.5</td>
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<td>78.30 ± 1.8</td>
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|                  |          |        |          |          |          |
|                  | a        | b      |          |          |          |
|                  | 81.1 ± 1.9 | 76.50 ± 2.1 | 86.14 ± 2.4 | 84.02 ± 2.7 | 88.26 ± 2.1 |
|                  |          |        |          |          |          |
|                  | 80.75 ± 2.4 |        |          |          |          |
|                  | 83.21 ± 2.3 |        |          |          |          |
|                  | 79.87 ± 1.9 |        |          |          |          |
|                  | 75.49 ± 2.4 |        |          |          |          |
|                  | 79.63 ± 1.6 |        |          |          |          |
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