Constraining the age and stability of unconsolidated deposits with cosmogenic nuclides in the McMurdo Dry Valleys, Antarctica

Daniel J. Morgan

A dissertation
submitted in partial fulfillment of the
requirements for the degree of

Doctor of Philosophy

University of Washington

2009

Program Authorized to Offer Degree:
Department of Earth and Space Sciences

University of Washington
Graduate School

This is to certify that I have examined this copy of a doctoral dissertation by

Daniel J. Morgan

and have found that it is complete and satisfactory in all respects,
and that any and all revisions required by the final
examining committee have been made.

Chair of the Supervisory Committee:

____

Jaakko Putkonen

Reading Committee:

____

Jaakko Putkonen

____

John O.H. Stone

____

Eric Steig

Date:____________________________

University of Washington

Abstract

Constraining the age and stability of unconsolidated deposits with cosmogenic nuclides in the McMurdo Dry Valleys, Antarctica

Daniel J. Morgan

Chair of the Supervisory Committee:
Associate Professor Jaakko Putkonen
Department of Earth and Space Sciences

The focus of this dissertation is the evolution of unconsolidated deposits in the McMurdo Dry Valleys, Antarctica (MDV) over millions of years. Using the concentration of cosmogenic nuclides ^{10}Be and ^{26}Al from vertical profiles in meter-deep hand-dug soil pits, I establish degradation rates and limiting ages for unconsolidated deposits in the MDV. By quantifying rates of geomorphic degradation that operate for millions of years in the MDV, I provide a clearer picture of how these landscapes are actively changing over long periods of time, which will allow us to better interpret the environmental information contained in unconsolidated deposits there.

The presumed antiquity of the landscapes of the MDV is based on the presence of ash deposits at or near the surface that are up to ~15 million years (Ma) old. For two of these ash deposits, the Arena Valley ash (4.33 Ma) and the Hart ash (3.9 Ma), I show that the measured concentrations of ^{10}Be and ^{26}Al are consistent with the ages of the ashes only if the regolith at these sites is steadily eroding at rates between 0.2 – 4.7 m Ma^{-1}. By sampling three glacial deposits in the MDV, I show that some glacial deposits indicate that they have been modified more by the sublimation of ice than by the erosion of the till. The degradation rates that best match the data at these three sites

range from 0.7 – 12 m Ma^{-1} for the sublimation of ice with initial debris concentrations ranging from 7.7 – 33% and erosion of the overlying till at rates of 0.5 – 1.2 m Ma^{-1}.

Collectively, these results show that the MDV are a more dynamic landscape than previously thought. At all of these sites, erosion appears to occur with extremely little to no vertical mixing or soil creep, which is a process that is ubiquitous on Earth. The low erosion rates determined in this study, coupled with the apparent lack of creep below the upper few centimeters of the surface may explain the apparent contradiction between landforms that maintain their meter-scale morphology for millions of years despite the fact that they are steadily eroding.

TABLE OF CONTENTS

LIST OF FIGURES

Figure Number | Page

LIST OF TABLES

ACKNOWLEDGEMENTS

There are many people to thank for helping me along this journey. First and foremost I have to thank my wife, Torrey, for her unconditional love and support throughout this entire process, especially through the many months that we spent apart while I did field work in Antarctica. I would also like to thank all of the Morgans and Sheltons for their support and encouragement while I was in graduate school.

I would like to thank my advisor, Jaakko Putkonen, for giving me the opportunity to work with him in one of the most amazing field areas in the world. My committee members John Stone and Eric Steig taught me the analytical and laboratory techniques that I needed to accomplish the goals set out in this project. Special thanks are due to Greg Balco who provided me with his infinite wisdom and mentored me through the process of getting a PhD. Joyce Cooper was a fantastic graduate school representative on my committee and provided dependable advice throughout my entire time at UW. Field assistance in the McMurdo Dry Valleys from Brendan O'Donnell, Nathan Turpen, Jon Connolly, and Kate Craig was greatly appreciated.

I want to thank all of the graduate students in the DOEASS for putting up with my shenanigans and supporting me throughout this process. My lasting memories from graduate school will be of our intramural ultimate frisbee championships, Liquefaction, Friday basketball games, and the numerous exploits that are not fit to print. Rock Stars, keep your nose to the grindstone and your eyes on the stars.

DEDICATION

To the woman who rated my buns AAA, Torrey.

Chapter 1

INTRODUCTION

Geomorphology is the study of the landscapes on the surface of the earth and the processes that alter these landforms through time, resulting in landscape evolution. Because the surface of the Earth is where most human activity occurs and landforms record information about the environment, understanding the processes that form and alter landscapes is important. Early models of landscape evolution proposed the idea that landforms would progress through a sequence of stages [e.g. *Davis*, 1899], or that landscapes would reach an equilibrium between form and erosion rate [e.g. *Gilbert*, 1909; *Hack*, 1975], whereas current research directions are investigating the interactions that the governing processes of climate, tectonics, and erosion have over landscape development [e.g. *Molnar and England*, 1990]. Although much progress has been made to advance our knowledge of surface processes, a major gap in our understanding is how unconsolidated deposits change over long periods of time. Unconsolidated deposits are the loose, non-cemented sediments that typically lie at or near the surface of the Earth [*Bates and Jackson*, 1984].

The recent developments in the methods that utilize cosmogenic nuclides to examine the exposure history of surficial deposits make it possible to study surface processes from $10^2 – 10^7$ years [*e.g. Lal*, 1991; *Gosse and Phillips*, 2001], which allows us to address long-standing questions regarding the longevity of landscapes over millions of years. Interpreting the paleoclimate information that is recorded by surficial landforms that are millions of years old is complicated by the degree to which these deposits have degraded since they were deposited. Thus, the advances in analytical techniques that allow us to measure extremely low rates of surface degradation with cosmogenic nuclides allow us to examine how landscapes evolve over millions of years and address fundamental questions regarding the interpretation of paleoclimate information that ancient landscapes and landforms contain.

In this dissertation, I address the issue of landscape evolution over long periods of time in the McMurdo Dry Valleys, Antarctica (MDV), which are thought to be among the oldest landscapes on Earth [*e.g. Denton et al.*, 1993]. Though previous research has addressed bedrock erosion over millions of years in arid regions [*e.g. Cockburn et al.*, 1999; *Nichols et al.*, 2006], very little work has examined how unconsolidated deposits degrade in arid regions like the MDV. Additionally, the MDV contain a large terrestrial archive of Antarctic Neogene glacial history in the form of glacial deposits and landforms, and understanding how the Antarctic Ice Sheet has fluctuated through time and responded to past changes in climate is important to the development of climate impact scenarios [*Intergovernmental Panel on Climate Change*, 2007]. Thus, it is fundamental to understand how unconsolidated deposits in the MDV evolve over millions of years so that we can better interpret the environmental information that they contain. I accomplish this by using cosmogenic nuclides ^{10}Be and ^{26}Al from vertical profiles dug into unconsolidated deposits, from which I quantify degradation rates and establish limiting ages of the landforms in the MDV.

1.1. Statement of research problem and goals

The McMurdo Dry Valleys of Antarctica contain a number of well-mapped glacial deposits that record numerous episodes of Neogene glaciation [*e.g. Calkin*, 1964; *Nichols*, 1966; *Hendy et al.*, 1979; *Denton et al.*, 1984; *Brook et al.*, 1993; *Hall et al.*, 1993; *Marchant et al.,* 1993a, 1993b; *Wilch et al.*, 1993; *Lewis et al.*, 2007]. These deposits record different types of glaciation, including the expansion of local alpine glaciers, the growth of regional outlet glaciers, and the overriding of mountain ranges by a northeast-flowing East Antarctic Ice Sheet. In order to correctly interpret the record of past glacial expansions in the MDV recorded by unconsolidated deposits, we need to understand how geomorphic processes that operate in the cold, polar and hyperarid Antarctic conditions have altered these deposits. The goals of this study are to use cosmogenic nuclides ^{10}Be and ^{26}Al that are produced *in situ* in quartz sands to establish limiting depositional ages and degradation rates for unconsolidated deposits in the MDV. This type of geomorphological analysis is the foundation for deciphering the record of former environments contained in unconsolidated deposits. In order to correctly interpret the archives contained in these deposits, we must first understand any changes that they have undergone since they were deposited.

Unconsolidated deposits in the MDV have been given absolute or limiting ages from cosmogenic nuclide exposure ages from boulders on moraines [e.g. *Brook et al.*, 1993; *Staiger et al.*, 2006] or by the stratigraphic relationships between deposits that contain dateable material such as ash deposits or basalt clasts [*e.g. Wilch et al.*, 1993; *Marchant et al.,* 1993a, 1993b; *Hall et al.*, 1993; *Lewis et al.*, 2006]. The preservation of ash deposits, the oldest of which range in age from 4 – 15 Ma, at the present-day surface has been interpreted as indicating that the climate in the MDV has remained cold and dry for millions of years and that little to no landscape degradation has occurred over this time [*e.g. Denton et al.*, 1993; *Marchant et al.*, 1996; *Lewis et al.*, 2007]. Supporting the notion that the MDV are a relict landscape where degradation rates are vanishingly small is the observation that many glacial moraines and landslide deposits have maintained their meter-scale morphology even though they are millions of years old. However, bedrock and regolith erosion rates in the MDV are 0.1 – 4 m Ma^{-1}, which suggests that the landscape should have been visibly transformed over a period of millions of years [*Summerfield et al.*, 1999; *Putkonen et al.*, 2008]. Chapter 2 explores the interpretation that ash deposits in the MDV indicate slope stability and the apparent contradiction between unconsolidated deposits that are millions of years old and have maintained their meter-scale morphology despite the fact that they are eroding at rates on the order of meters per millions of years.

Another intriguing feature of glacial deposits in the MDV is that ground ice persists near the surface in Miocene-aged soils in the MDV even though modeled and measured sublimation rates of ice in soils suggest that, without any recharge mechanisms, ground ice should sublimate at rates that would result in the upper few meters of soil being ice-free on the order of 10^3 – 10^5 years [*Hindmarsh et al.*, 1998;

McKay et al., 1998; *Campbell and Claridge*, 2006; *Bockheim et al.*, 2007; *Hagedorn et al.*, 2007]. The potential for ice to persist in the MDV, and the argument surrounding its age, is best illustrated by the buried glacial ice that underlies the Granite Drift in central Beacon Valley that is thought to be as young as 310 ka [*Ng et al.*, 2005] and perhaps as old as 8.1 Ma [*Sugden et al.*, 1995; *Marchant et al.*, 2002]. There is considerable ongoing debate regarding the age and stability of ground ice in the MDV, and Chapter 3 explores the degradation of glacial deposits in the MDV due to both the sublimation of ice and the erosion of till.

The primary goals of this project are to quantify degradation rates and establish limiting ages for unconsolidated deposits in the MDV. At a large scale, this work is motivated by the lack of understanding of how landscapes will evolve over millions of years under the unique, cold, polar, and hyperarid conditions that are present in the MDV today. At a smaller scale, this project is further motivated by the apparent conflict between the persistence of landforms that are eroding at rates on the order of ~m Ma^{-1} while maintaining their meter-scale morphology, and the presence of ground ice within Miocene-aged soils in the MDV. To accomplish these goals, I use concentrations of cosmogenic nuclides ^{10}Be and ^{26}Al in vertical profiles from hand dug soil pits to study how quickly and by what processes unconsolidated deposits in the MDV degrade. Once we understand the degree to which these deposits have been altered since they were deposited, then we can correctly interpret the environmental information that they record.

1.2. Field Area

The McMurdo Dry Valleys are the largest ice-free area in the Transantarctic Mountains [*Denton et al.*, 1993] (Figure 2.1). The MDV are free of ice because sublimation rates are equal to or higher than precipitation rates [*MacClune et al.*, 2003] and because glacial ice from the Taylor dome is diverted around the MDV and through the Ferrar and Mackay Glaciers. Local bedrock in the area consists of Devonian-Triassic aged sedimentary rocks (Beacon Supergroup), which overlie a Precambrian complex of metamorphic basement rocks, and are intruded by the Jurassic aged dolerite in the form of sills and dikes (Ferrar Supergroup) [*Barrett*, 1981], and there are numerous Cenozoic volcanoes within and adjacent to the MDV [*Kyle*, 1990]. Denton et al. [1993] propose that the MDV were initially formed by fluvial action in a semi-arid environment similar to the desert southwest of North America and have since been modified by glacial action.

Mean annual temperature and precipitation in the MDV are controlled by elevation, distance from the coast, and exposure to katabatic winds, resulting in a strong east-west climate gradient [*Fountain et al.*, 1999; *Doran et al.*, 2002]. The mean annual temperature on the valley floors of Wright and Taylor Valleys is between -17.7°C and -27.4°C [*Doran et al.*, 2002] and -24°C in Beacon Valley [*Putkonen et al.*, 2003]. The annual precipitation, as measured at Lake Vanda (93 masl in Wright Valley), is less than 100 mm water equivalent [*Fountain et al.*, 1999]. Though running melt water is common during the austral summer in the lower valleys (~200 masl), it is rarely observed in the upper valleys (> 1000 masl). Many researchers have classified microclimate zones in the MDV [*e.g. Campbell and Claridge*, 1969, 1987, 2006; *Bockheim*, 2002]. Marchant and Denton [1996], defined microclimate zones by the characteristic equilibrium geomorphic landforms that would form in each microclimate. Marchant and Head [2007] expanded upon these definitions and called them the stable upland zone, the inland mixed zone, and the coastal thaw zone.

Under the cold, polar, hyperarid climate conditions that exist in the MDV today, erosion rates are thought to be extremely low. Based on the presence of ash deposits at or near the surface, it has been proposed that the landscapes of the MDV have been exposed sub-aerially since the mid-Miocene, and that essentially zero landscape evolution has occurred since then [*e.g. Marchant et al.*, 1996]. Using the concentration of cosmogenic nuclides ^{10}Be and ^{26}Al in unconsolidated deposits we can establish degradation rates for these deposits and directly test the antiquity of these deposits.

1.3. Geomorphic processes active in the MDV

This section will briefly review what is known about the geomorphic processes that are actively transforming the landscape of the MDV today. Because running water is only active in a few lowland areas in the MDV and is not active at any of the sample sites, I will not consider the effects that it has on the landscape of the MDV. Ignoring running water, the main geomorphic processes that are active in the MDV today and that affect the morphology are katabatic winds, active-layer cryoturbation, and local alpine glaciation.

The MDV are known for strong winds, especially the gravity driven katabatic winds [*Schwerdtfeger*, 1984], and a variety of erosional and depositional eolian landforms are found throughout the MDV [*Morris et al.*, 1972]. At Beacon Valley, mean daily wind speed over a three-year period was 4.2 m s^{-1}, the mean of maximum daily 15 minute wind speeds was 7.6 m s^{-1}, and the max 15 minute wind speed was 22 m s^{-1} [J. Putkonen, personal communication, 2009]. In Victoria Valley, the average wind speed recorded over a three-year period was 5 m s^{-1}, and the katabatic wind speeds averaged 40 m s^{-1} and reached a maximum of 64 m s^{-1} [*Hagedorn et al.*, 2007]. By installing sediment traps ~1m above the surface, Lancaster [2002] determined the total eolian flux of sediment in the MDV varied from 0.26 g m^{-2} yr^{-1} to 441.8 g m^{-2} yr^{-1}. Not only do strong winds transport material, but they also redistribute significant amounts of snow throughout the MDV, and the distribution of this snow can greatly affect the microclimate of unconsolidated deposits.

Typically, the active layer is defined as the surface horizon above permafrost that experiences seasonal temperature fluctuations above and below 0°C [*Washburn*, 1980]. In the MDV, it is necessary to distinguish between both "wet" and "dry" active layers and permafrost because many soils in the MDV lack sufficient soil moisture to contain ice and/or liquid water [*Bockheim et al.*, 2007]. In general, the distinction between "wet" and "dry" permafrost and active layers is at 5% gravimetric water content, which is generally taken as the necessary amount of moisture necessary to activate frost heave [*Marchant and Head*, 2007]. In the Sør Rondane Mountains, when the regolith moisture content exceeded 5%, daily frost heave cycles up to 3.4 mm were recorded, which contributed to the down slope movement of regolith at a rate of 0.015 m yr^{-1} [*Matsuoka and Moriwaki*, 1992]. Contraction cracks in patterned ground terrain in the MDV spread at rates from 0.1 mm yr^{-1} up to 1.8 mm yr^{-1}, which has the potential to fully alter a surface in 10^3 to 10^6 years [*Sletten et al.*, 2003].

The sublimation rate of ice through regolith in the MDV has been directly measured, calculated from cosmogenic isotope profiles, and modeled from atmospheric principles. The present-day sublimation rates of bare ice in the MDV are on the order of $10^4 – 10^5$ m Ma^{-1} [*Clow et al.*, 1988]. At a site on the Linnaeus Terrace in Wright Valley (1600 masl), McKay et al. [1998] measured soil temperatures and found that the ice-cemented permafrost that began 25 cm below the surface was subliming at a rate of 0.4 – 0.6 mm yr^{-1}. At the site in Beacon Valley where the Granite Drift overlies massive relict glacier ice, Ng et al. [2005] calculated minimum sublimation rates as low as 0.004 mm yr^{-1} from cosmogenic nuclide concentrations from clasts in the overlying till. Stone et al. [2000] calculated sublimation rates from concentrations of cosmogenic nuclides from paired samples of the debris in the ice and the ice itself, and found that the long-term average sublimation rate was 0.05 mm yr^{-1}. Hindmarsh et al [1998] modeled ice sublimation through the ablation till with an atmospherically-determined vapor-pressure gradient (AVPG) model and found that the upper few meters of the till should be free of ice on the order of $10^3 – 10^5$ years. Without any recharge mechanisms, all of these rates suggest that the upper few meters of millions of years old soils in the MDV should be ice-free, but ice is found in the form of ground ice, buried ice, or ice-cemented permafrost in many of the Miocene and Pliocene-aged soils in the MDV.

Alpine glaciers are common in the MDV, and all of them are cold-based [*Chinn*, 1980], meaning that they are below the pressure melting point at their base. Because cold-based glaciers are frozen to their bed they are thought to be only weakly erosive. Cuffey et al. [2000] found that even though the Meserve Glacier in Wright Valley has a basal temperature of -17°C, it does actively entrain material at subfreezing temperatures and may have created the u-shaped trough that the glacier occupies. Cuffey et al. [2000] calculated a bedrock erosion rate of ~$1 - 3 \times 10^{-7}$ m yr^{-1} for the Meserve Glacier, which is much lower than for typical temperate glaciers [*Hallet et al.*, 1996]. Because this erosion rate is lower than most measurements of bedrock erosion rates in the MDV [*Summerfield et al.*, 1999], it is suggested that cold-based glaciers in the MDV are actually protective of the underlying landscapes [*Marchant and Head*, 2007].

By measuring the concentration of ^{10}Be and ^{26}Al in soil profiles of unconsolidated deposits in the MDV I will be able to test for the activity of and quantify the rates of certain geomorphic processes (e.g. stability, erosion, soil creep, or burial). The general concept for analyzing the cosmogenic nuclide measurements is to use the local geologic and geomorphic context to determine what happened geologically, and then use the concentration of ^{10}Be and ^{26}Al to determine when and how fast the geologic events occurred. I accomplish this by constructing an exposure model that describes the geologic events and predicts the concentration of ^{10}Be and ^{26}Al based on certain parameters like the exposure duration, erosion rate, sublimation rate of ground ice, vertical mixing rate, or shielding from exposure to cosmic rays. By optimizing the parameters in the exposure model to best match the measured nuclide predictions, I can constrain the rates at which certain geomorphic processes are active in the MDV.

1.4. Glacial history of the MDV

Understanding how the Antarctic Ice Sheet has fluctuated through time and responded to past changes in climate is critical to the development of climate impact scenarios [*Intergovernmental Panel on Climate Change*, 2007]. While it is generally accepted that the marine-based West Antarctic Ice Sheet (WAIS) has experienced significant volumetric changes in the past few million years [*e.g. Naish et al.*, 2009], the history of the much larger and terrestrial-based East Antarctic Ice Sheet (EAIS) is the subject of ongoing debate [e.g. *van der Wateren and Hindmarsh*, 1995]. Because the MDV are located in a part of the Transantarctic Mountains that are thought to have been overridden by the EAIS, they are bounded by the Ferrar and MacKay glaciers that drain the EAIS, and they are the largest ice-free area on the continent, unconsolidated deposits in the MDV are central to our understanding of the evolution of the EAIS. The main unresolved issue regarding the history of the EAIS is when it switched from a warm-based ice sheet to the present cold-based one. Arguments for when this change occurred generally fall into two camps, the “stable hypothesis” that makes the case that the EAIS reached its current size and environmental state ~15 Ma and has remained this way since then [*e.g. Sugden et al.*, 1993], and the “dynamic hypothesis” that contends that the EAIS remained a warmer, wet-based ice sheet that underwent significant fluctuations in volume and extent during the period from 3 – 15 Ma ago and did not reach its current size and climatic state until the late Pliocene [*e.g. Webb et al.*, 1984; *Barrett et al.*, 1992].

Much of the evidence supporting the stable hypothesis comes from the MDV where unweathered *in situ* ash deposits with ages as old as 4 – 15 Ma lie near the surface [*e.g. Marchant et al.,* 1993a, 1993b; *Marchant et al.*, 1996; *Lewis et al.*, 2006]. Lewis et al. [2007], report finding 14 Ma old lake deposits in the Olympus Range that they interpret as the last deposits from the alpine environment that existed before the current cold and hyperarid climate took hold. The climatic interpretation of these deposits hinges on their pristine preservation, which, the authors argue, precludes a warmer or wetter climate since these units were deposited, otherwise the ashes would have weathered to clays and the organic material in the lake deposits would have decayed. Some data from ocean sediment cores support the idea that the EAIS has been stable since the mid-Miocene and that it did not experience large fluctuations that the dynamist viewpoint supports. Based on oxygen isotope data Kennett and Hodell [1995] conclude that Pliocene sea surface temperatures were not different from that today and Warnke et al. [1996] show that ice-rafted debris concentrations did not decrease during the Pliocene, which suggests that the EAIS was intact and stable during this period.

The evidence supporting the contrasting dynamic hypothesis of the history of the EAIS comes mainly from the Sirius Group deposits found throughout the Transantarctic Mountains and from ocean sediment cores. The Sirius Group deposits are a mixture of diamicts that record glacial, fluvioglacial, glacial-marine, and lacustrine environments that are thought to be Pliocene in age based on the inclusion of reworked marine diatoms [*Webb et al.*, 1984; *Barrett et al.*, 1992]. Because the Sirius Group deposits are found at high elevation in the Transantarctic Mountains (> 2,000 masl), the dynamic interpretation of the Sirius Group is that the EAIS was first substantially smaller than it was today during the Pliocene such that open seaways were common in the Antarctic interior, and that the EAIS subsequently was much larger than it was today such that it overrode the Transantarctic Mountains and deposited the Sirius Group deposits at the high elevations where they are found today. Though there is significant evidence that shows that marine diatoms are easily transported by eolian process and are found even at the South Pole [*Kellogg and Kellogg*, 1996; *Stroeven et al.*, 1996], Harwood and Webb [1998] argue that these diatoms were transported by glacial processes. Ocean sediment records from the CIROS-2 core in the Ferrar Fjord show repeated cycles of glacial advance and retreat from the period 2.0 – 4.9 Ma, which are interpreted as changes in the EAIS and not incursions of the WAIS [*Barrett and Hambrey*, 1992]. The Pagodroma Group found in the Lambert Graben region of the EAIS are Miocene and Pliocene-aged fjordal deposits that are interpreted as indications of major EAIS retreat because they are found 250 – 300 km inland from the current ice-ocean margin [*McKelvey et al.*, 2004]. Taken together, this evidence suggests that during the Pliocene the EAIS has fluctuated between being significantly smaller and much larger than it is today, and that it did not reach the current cold hyperarid climate that it has today until the late Pliocene.

The stable and dynamic hypotheses may appear to be incompatible, but an emerging view is that the MDV have remained cold and hyperarid since the Miocene even though the EAIS and WAIS have experienced major fluctuations [*Fox*, 2008]. Though the goals of this project are geomorphologic, the results will be directly related to the interpretations of the glacial history of the EAIS. Central to the stable EAIS hypothesis is the notion that Miocene-aged deposits in the MDV have persisted with minimal disturbance since they were laid down. By examining the stability of unconsolidated deposits in the MDV I can directly test the notion that these deposits have remained undisturbed since they were deposited. Determining the degree to which unconsolidated deposits have degraded since they were deposited is fundamental to making the correct paleoclimate interpretation based on these sediments.

1.5. Organization of this dissertation

In Chapter 2, I explore the interpretation that ash layers in the MDV indicate slope stability by measuring the concentration of ^{10}Be and ^{26}Al from colluvium deposits below two well-studied ash layers, the Arena Valley ash (^{40}Ar/^{39}Ar age of 4.33 Ma) and the Hart ash (K-Ar age of 3.9 Ma)., as well as from sands intermixed with the upper layer of the ashes. At both of the ash sites, the measured concentrations of ^{10}Be and ^{26}Al are significantly lower than expected given the age of the *in situ* airfall ashes, and I investigate the possibility that burial by ice or erosion of the regolith can explain these low nuclide concentrations. In Chapter 3, I address the apparent contradiction between millions of years old glacial deposits that contain ground ice and maintain their meter-scale morphology despite the fact that measured and modeled degradation rates suggest that these deposits should be ice-free and have noticeably eroded over these time periods. By analyzing the concentration of cosmogenic ^{10}Be and ^{26}Al in soil profiles from three glacial deposits in the Quartermain Mountains, Antarctica, I found that the measured nuclide concentrations are inconsistent with the stratigraphic ages of the tills and that at some sites the sublimation of ice and erosion of the overlying till can best explain the measured nuclide concentrations. Chapter 4 discusses the conclusions that can be drawn from the results of this work, and poses questions for future work.

Chapter 2

REGOLITH EROSION RATES
IN THE MCMURDO DRY VALLEYS, ANTARCTICA

2.1. Introduction

The McMurdo Dry Valleys (MDV) in Antarctica are the largest ice-free area on the continent [*Drewry*, 1983], and are among the oldest landscapes on Earth [*e.g. Denton et al.*, 1993; *Schafer et al.*, 1999]. The MDV also contain a number of well mapped glacial deposits that record various episodes of Neogene glaciation [*Taylor*, 1914; *Calkin*, 1964; *Nichols*, 1966; *Denton et al.*, 1984; *Hall et al.*, 1993; *Marchant et al.,* 1993a*,* 1993b]. In order to correctly interpret the record of past glacial expansions in the MDV recorded by these deposits we need to understand how geomorphic processes that operate in the cold and dry Antarctic conditions have altered them. Additionally, very little is known about the degradation of unconsolidated deposits in ancient landscapes that appear unchanged for millions of years. Once we understand how the deposits may have been altered after deposition, then we can reconstruct the age and paleoclimatic information that they may contain.

In this paper we use the term regolith to mean the loose, incoherent mantle of rock fragments, soil, glacial drift, blown sand, etc. that rests upon solid bedrock [*Whitten and Brooks*, 1987]. We will use this term in this paper when the source of unconsolidated material in the MDV is not known, and therefore terms like colluvium or eolian deposit may not be appropriate. We will use the term colluvium to describe units that have been defined as such by previous researchers.

Our general understanding of regolith degradation is based on studies of Holocene- and ice-aged scarps, hillslopes, and moraines from the mid latitudes [*Nash*, 1980; *Hanks et al.*, 1984; *Fernandes and Dietrich*, 1997; *Hanks*, 2000; *Putkonen and O'Neal*, 2006]. In the MDV, present-day regolith movement has been observed [*Putkonen et al.*, 2007] and eolian sediment fluxes have been measured [*Lancaster*, 2002], but very few measurements of regolith degradation have been made [*Putkonen* *et al.*, 2008*; Schiller et al.*, 2009]. The erosion rate of the exposed bedrock in the MDV is weathering limited, meaning that once material is broken out of the bedrock, it is efficiently removed by transport processes and leaves the bedrock surface free of debris [*Selby*, 1971]. We expect that the erosion rate of regolith in the MDV would be equal to or greater than the rate of erosion of bedrock, which ranges from 0.133 – 1.02 m Ma^{-1} in the MDV [*Summerfield et al.*, 1999].

Unconsolidated deposits in the MDV have been given absolute or limiting ages from cosmogenic nuclide exposure ages of boulders on moraines [*Brook et al.*, 1993, 1995; *Staiger et al.*, 2006] or by their stratigraphic relationships to deposits that contain dateable material such as ash deposits or basalt clasts [*Wilch et al.*, 1993; *Marchant et al.,* 1993a, 1993b; *Hall et al.*, 1993; *Lewis et al.*, 2006]. The preservation of ash deposits up to 15 Ma old at the present-day surface has been interpreted as indicating that the climate in the MDV has remained cold and dry for millions of years and that little to no landscape evolution has occurred over this time [*Denton et al.*, 1993; *Hall et al.*, 1993; *Marchant et al.,* 1993a, 1993b; *Marchant and Denton*, 1996; *Marchant et al.*, 1996; *Lewis et al.*, 2007; *Marchant and Head*, 2007]. Supporting the notion that the MDV are a relict landscape where degradation rates are vanishingly small is the observation that many glacial moraines and landslide deposits have maintained their meter-scale morphology even though they are millions of years old. Additional motivation comes from the fact that very little is known about how unconsolidated deposits degrade in the unique environment that the MDV encompass.

To explore the apparent contradiction between the measured bedrock and regolith rates that range from 0.1 – 4 m Ma^{-1}, [*Summerfield et al.*, 1999; *Putkonen et al.*, 2008] and the implied slope stability by the presence of millions of years old *in situ* deposits [*Hall et al.*, 1993; *Marchant et al.,* 1993a; *Marchant et al.*, 1996], we concentrate on two well documented air-fall ash deposits. These are the Arena Valley ash in Arena Valley and the Hart ash in Wright Valley (Figure 2.1). Although the upper parts of these ashes show some mixing of non-volcanic material with the ash, both ashes are interpreted as primary deposits because of sharp basal contacts with the underlying colluvium, and their basal layers lack non-volcanic contaminants. The Arena Ash has a $^{40}Ar/^{39}Ar$ age of 4.33 ± 0.07 Ma [*Marchant et al.*, 1993a] and the Hart ash has a K-Ar age of 3.9 ± 0.3 Ma [*B.L. Hall et al.*, 1993]. The ages of these ashes are generally accepted and have been used to develop relative ages of other units that have a stratigraphic relationship with them [*e.g. Hall et al.*, 1997; *Prentice and Krusic*, 2005].

The objective of this study is to quantify regolith degradation rates in a cold polar desert. Using the concentration of cosmogenic nuclides ^{10}Be and ^{26}Al in quartz sands in unconsolidated deposits, we have examined the post-depositional geologic history of these deposits, including the exposure duration, erosion rate, and burial history. This has allowed us to investigate the apparent contradiction between the preservation of meter-scale features for up to 15 Ma and erosion rates at the ~m Ma^{-1} scale. Furthermore, determining the rates at which regolith degrades in the MDV will advance our understanding of landscape evolution in cold and hyperarid environment.

2.2. Field area and sample sites

Situated between the East Antarctic Ice Sheet and the Ross Sea (Figure 2.1), the MDV are free of ice because ice sublimation rates are equal to or higher than snow precipitation rates [*MacClune et al.*, 2003] and because glacial ice from the Taylor dome is diverted around the MDV and through the Ferrar and Mackay Glaciers [*Chinn*, 1990]. Mean annual temperature and precipitation in the MDV are controlled by elevation, distance from the coast, and exposure to katabatic winds, resulting in a strong east-west climate gradient [*Fountain et al.*, 1999; *Doran et al.*, 2002]. Many researchers have classified microclimate zones in the MDV [*Campbell and Claridge*, 1969, 1987, 2006; *Bockheim*, 2002]. Marchant and Denton [1996], and later expanded by Marchant and Head [2007] defined these zones by the characteristic equilibrium geomorphic landforms that would form in each microclimate. These zones are the stable upland zone, the inland mixed zone, and the coastal thaw zone. The Arena

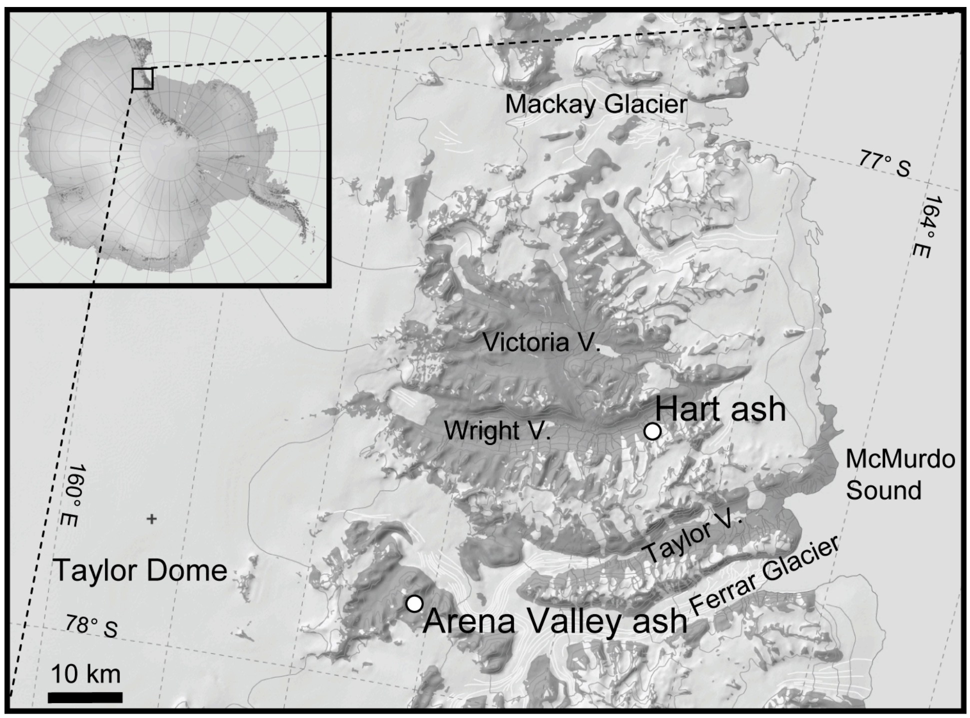

Figure 2.1. Map of Antarctica (inset) and the McMurdo Dry Valleys. The white circles show the location of the two ash deposits studied in this chapter.

Valley ash is found in the stable upland zone (mean annual temperature -22°C) and the Hart ash is found in the inland mixed zone (mean annual temperature -18°C). The mean annual temperature on the valley floor of Wright Valley is -17.7°C [*Doran et al.*, 2002], and in Beacon Valley (which is adjacent to Arena Valley) it is -24°C [*Putkonen et al.*, 2003]. The annual precipitation, as measured at Lake Vanda (93 masl in Wright Valley), is less than 100 mm water equivalent [*Fountain et al.*, 1999]. Though running melt water is common in the lower valleys (~200 masl) during the austral summer, it is rarely observed in the upper valleys (> 1000 masl).

The Arena Valley ash was first described by Marchant et al. [1993c], and was later described in more detail by Marchant et al. [1993a]. As described by Marchant et al. [1993a], the ash is about 25 cm thick, underlies the modern desert pavement, and was deposited on a desert pavement that armors the Monastery colluvium. Marchant et al. [1993a] described the deposit as having a basal unit that is 0.5-1.0 cm thick that lack non-volcanic contaminants and lies in sharp stratigraphic contact with the underlying desert pavement. Based on the sharp stratigraphic contact with the underlying desert pavement and the lack of non-volcanic contaminants, Marchant et al. [1993a] interpreted the Arena Valley ash as an *in-situ* airfall ash. The $^{40}Ar/^{39}Ar$ age of sanidine in the ash is 4.33 ± 0.07 Ma.

Following the description of the site in Marchant et al. [1993a], we found the Arena Valley ash near the foot of the slope of the main tributary valley in west-central Arena Valley (04-AV-Pit 4: 77.86378°S, 160.84510°E, 1,364 m). We measured depth down from the surface beginning at the bottom of the modern desert pavement and found that 0-3 cm was a mixture of colluvial sands and some ash, 3-18 cm was entirely ash, 18-20 cm was the buried desert pavement, and below 20 cm was the Monastery colluvium (Figure 2.2). Visual inspection of the basal unit of the ash under magnification showed that delicate glass shards and spires remained intact, suggesting that the ash was an airfall ash.

The Hart ash in Lower Wright Valley was first described by Hall et al [1993] and was found at three different localities. As described by Hall et al. [1993], the ash is

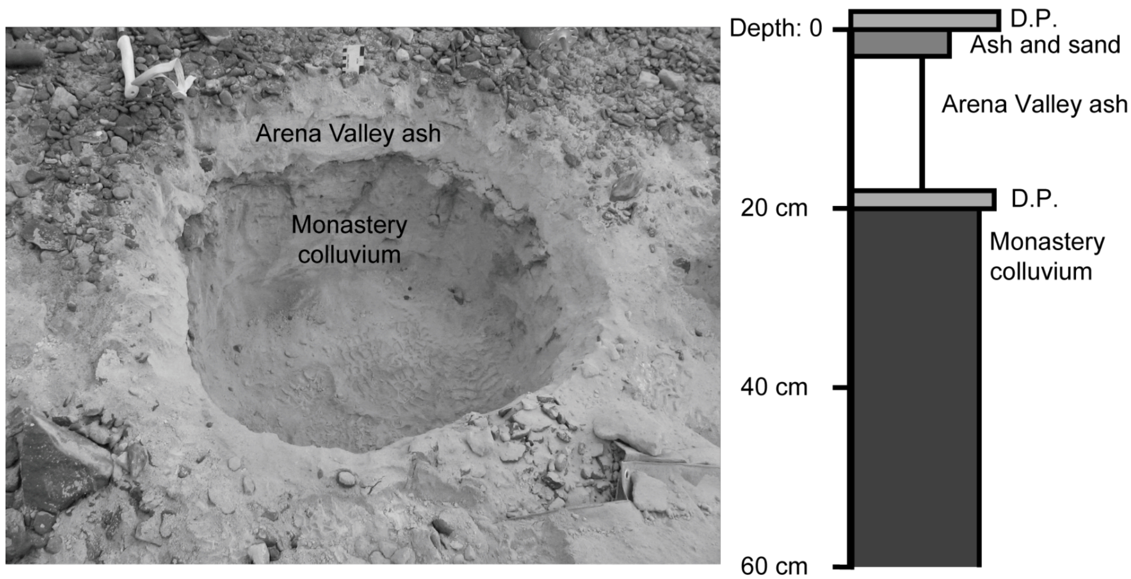

Figure 2.2. The Arena Valley ash and the stratigraphy of the site (04-AV-Pit 4). Widths of the stratigraphic columns represent relatively larger average sediment grain size of the unit. D.P. indicates a desert pavement. At this site, the Arena Valley ash is 15 cm thick, beginning 3 cm below the desert pavement that makes the current surface. The Arena Valley ash has an $^{40}Ar/^{39}Ar$ age of 4.33 Ma (Marchant et al. 1993a) and rests on a desert pavement that caps the Monastery colluvium.

usually 30-40 cm thick, the upper 20-30 cm are usually disturbed and contain a few sand stringers and/or clasts, but the lower units are pure ash, undisturbed, and have a sharp basal contact with the underlying colluvium. In some excavations, Hall et al. [1993] found that the ash was deposited on a poorly developed desert pavement. Because the thinner edges of the ash deposits have been reworked, Hall et al. [1993] suggest that the ash may have been overridden by westward flowing ice after its deposition. Based on the sharp basal contact with the underlying colluvium and the undisturbed basal layers of the ash, Hall et al. [1993] interpreted the ash as a primary deposit, and glass in the ash has a K-Ar age of 3.9 ± 0.3 Ma. The largest deposit of the Hart ash (150 m x 85 m) found by Hall et al. [1993] was located between the Hart and Goodspeed glaciers at an elevation of 378 m.

Following the description of the most extensive deposit in Hall et al. [1993], we found the Hart ash between the Hart and Goodspeed glaciers in Lower Wright Valley (04-LWV-Pit 25: 77.49587°S, 162.37390°E, 386 m). The sample site (Figure 2.3) is capped by a gravel lag that consists of gravel and coarse sized sand that is mobile during high wind events, as evidenced by the gravel ripples that form in the surrounding area. The ash generally began 1-2 cm below this gravel lag, but in places a thin (< 10 cm) colluvium layer was above the ash. Where we sampled the sediment we found that the upper two cm of the soil were a mixture of sand and ash, and that pure ash began two cm below the gravel lag and reached a depth of 40 cm. Below the ash was a strongly discolored (most likely oxidized) colluvium layer. At this site we found a relict sand wedge that terminated abruptly at the colluvium-ash contact suggesting that freeze thaw activity predated the deposition of the ash. Additionally, we found a few sand stringers that cut into the ash suggesting that some soft sediment deformation occurred during the deposition of the ash. For cosmogenic nuclide analysis, we used samples from the colluvium below the ash and not the sand wedge or the sand stringers. Visual inspection of the basal unit of the ash under magnification found delicate glass shards and spires, suggesting that the ash was an airfall ash.

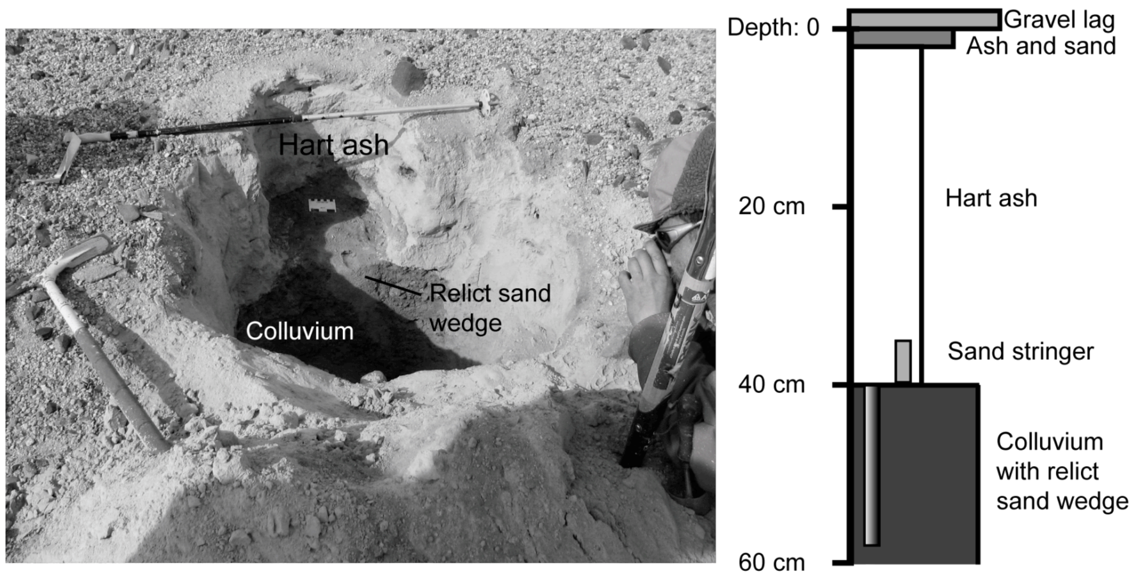

Figure 2.3. The Hart ash and the stratigraphy of the site (04-LWV-Pit 25). Widths of the stratigraphic columns represent relatively larger average grain size of the unit. At this site, the Hart ash is 38 cm thick, beginning 2 cm below the gravel lag that makes the current surface. The Hart ash has a K-Ar age of 3.9 Ma (Hall et al. 1993) and rests on a colluvium layer that contains a relict sand wedge. The sand wedge ends abruptly at the contact between the ash and the colluvium and does not penetrate the ash, indicating that any cryoturbation that formed the sand wedge pre-dates the deposition of the ash. Sand stringers do extend into the ash a few centimeters, indicating that the ash and colluvium underwent some soft-sediment deformation.

2.3. Methods

2.3.1. Sample collection and processing

In the field, we collected a series of bulk sediment samples from ~1-meter-deep hand-dug soil pits. At both sample sites we collected a sediment sample from just below the modern desert pavement, a series of samples of the ash itself, and a series of samples from the colluvium that underlies the ash. We measured the density of the colluvium and the ash by packing the sediment samples into a container of known volume. The sample depths and densities are found in Table 1.

To extract quartz from the sediment, we dry sieved the samples and sub-sampled the 0.3-0.5 mm size fraction. To remove pyroxene derived from the local dolerite, we used a heavy liquid separation, and the quartz was purified by repeated etching in 2% HF. We extracted beryllium and aluminum using standard methods [*Stone*, 2004] and measured $^{10}Be/^{9}Be$ and $^{27}Al/^{26}Al$ isotope ratios at the PRIME lab at Purdue University in West Lafayette, IN. The combined carrier and process blanks contained $1.67 \pm 0.462 \times 10^5$ atoms ^{10}Be and $0 \pm 2.04 \times 10^5$ atoms ^{26}Al. These are always less than 0.9% of the measured ^{10}Be atoms and 0.45% of the measured ^{26}Al atoms. The beryllium measurements were originally standardized to the ICN standard, and we have re-standardized them to the 07KNSTD [*Nishiizumi et al.*, 2007]. For aluminum, the isotope ratios were referenced to the Nishiizumi 2004 standard [2004]. We determined the ^{10}Be and ^{26}Al production rates at our sites by using the rates of Stone [2000]. ^{10}Be production rates were further reduced by a factor of 1.106 following Nishiizumi et al. [2007]. For the Arena Valley ash (04-AV-Pit 4) the topographic shielding factor for the site is 0.98 and the resulting surface production rates are 19.4 atoms g^{-1} yr^{-1} for ^{10}Be and 131.2 atoms g^{-1} yr^{-1} for ^{26}Al. For the Hart ash (04-LWV-Pit 25) the topographic shielding factor for the site is 0.99 and the resulting surface production rates are 8.21 atoms g^{-1} yr^{-1} for ^{10}Be and 55.7 atoms g^{-1} yr^{-1} for ^{26}Al. The nuclide concentration data are given in Table 1.

Table 2.1. Sample and isotope data. 'n.m' indicates that no measurements are available for this sample.

04-AV-Pit 4: Arena Valley ash					
Sample ID	Depth in soil (cm)	Soil density ($g \cdot cm^{-3}$)	Effective shielding mass ($g \cdot cm^{-2}$)	$^{10}Be \cdot 10^6 \pm$ 1 std uncertainty ($atoms \cdot g^{-1}_{quartz}$)	$^{26}Al \cdot 10^6 \pm$ 1 std uncertainty ($atoms \cdot g^{-1}_{quartz}$)
04-AV-Pit 4 0-1	0-1	1.78	0.907	11.4 ± 0.322	61.6 ± 2.07
Arena Valley ash	3-18	1.02	n.m.	n.m	n.m
04-AV-Pit 4 20-23	20-23	1.84	27.0	23.7 ± 0.994	88.6 ± 4.34
04-AV-Pit 4 30-33	30-33	1.83	45.1	20.6 ± 0.970	79.4 ± 2.59
04-AV-Pit 4 40-43	40-43	1.83	63.3	n.m.	74.6 ± 2.82
04-AV-Pit 4 50-53	50-53	1.80	81.4	15.0 ± 0.477	64.5 ± 2.80
04-AV-Pit 4 60-63	60-63	1.81	99.6	16.3 ± 1.41	54.7 ± 2.72
04-LWV-Pit25: Hart ash					
04-LWV-Pit 25 0-2	0-2	1.32	1.31	n.m.	15.4 ± 0.835
Hart ash	2-40	1.01	n.m.	n.m.	n.m.
04-LWV-Pit 25 45-48	45-48	1.76	52.9	2.60 ± 0.0757	10.5 ± 0.526
04-LWV-Pit 25 55-58	55-58	1.76	71.0	2.14 ± 0.0815	7.80 ± 0.406
04-LWV-Pit 25 65-68	65-68	1.81	89.1	2.16 ± 0.110	n.m.
04-LWV-Pit 25 77-80	77-80	1.91	111	1.99 ± 0.0948	5.72 ± 0.567

2.3.2. Exposure model

By using the geologic context to guide our interpretation of the cosmogenic nuclide concentrations we built an exposure model to explore the possible degradation of the regolith both before and after the ash is deposited. An advantage to measuring the nuclide concentrations at multiple depths is that we can examine specific questions about the nature of this degradation. The regolith could have remained undisturbed since the deposition of the ash, it could have been subjected to surface degradation, vertical mixing of the deposit due to an activity like cryoturbation, downslope creep processes, and any combination of these processes. Additionally, the sites could have been covered by weakly-erosive, cold-based ice sometime after the ash was deposited which would have strongly affected the resulting cosmogenic isotope concentrations of the underlying regolith.

The general concept for analyzing the cosmogenic nuclide measurements is to use the local geologic and geomorphic context to determine what happened geologically. We construct an exposure model to describe these geologic events and use the ^{10}Be and ^{26}Al concentrations to determine when and how fast the geologic events occurred. At both sample sites the general geomorphic context is similar, and the relevant aspects are as follows. Before either ash was deposited, a colluvium deposit (as termed by previous researchers) was exposed at the surface at each site. Then, ash was deposited directly on both of these colluviums. The presence of delicate glass shards and the lack of stratification in both ashes suggest rapid emplacement of a primary airfall ash. Finally, a desert pavement or gravel lag that formed at some point after the ash was deposited, caps each ash deposit.

The production of ^{10}Be and ^{26}Al at less than 1 m depth is almost entirely by spallation [*Gosse and Phillips*, 2001], which means that the production rate of ^{10}Be and ^{26}Al below the surface can be described by:

$$P_i(z) = P_i(0) \cdot e^{-z\ /\Lambda}, \tag{1}$$

where the subscript i indicates the nuclide of interest (either ^{10}Be or ^{26}Al), $P_i(0)$ is the surface production rate of nuclide i and has units of atoms g^{-1} yr^{-1}, z is the effective shielding mass of the sample and has units of g cm^{-2}, and Λ is the attenuation length, which is taken to be 150 g cm^{-2} in Antarctica (see Lal [1991] and Gosse and Phillips [2001] for a discussion of Λ). The shielding mass is the shielding provided by the overlying sediments for each sample, which is related to the depth of the sample below the surface and cumulative density of the overlying material. The effective shielding mass is the point in the sample where the isotope production rate (for either ^{10}Be or ^{26}Al) is equal to the average isotope production rate through the thickness of the sample.

The production and decay of nuclide i in sample j is governed by [*Lal*, 1991]:

$$\frac{dN_{i,j}}{dt} = P_i(z_j(t)) - N_{i,j}\lambda_i \quad , \qquad (2)$$

where N is the nuclide concentration that we have measured, P is the production rate at the sample site and is dependent on the shielding mass, z (g cm^{-2}), of the sample, and λ is the decay constant (yr^{-1}) for nuclide i, and the subscript j indicates the sample number. In this study we have used decay constants of $\lambda_{10} = 5.1 \times 10^{-7}$ yr^{-1} for ^{10}Be [*Nishiizumi et al.*, 2007] and $\lambda_{26} = 9.78 \times 10^{-7}$ yr^{-1} for ^{26}Al [*Nishiizumi*, 2004].

An exposure model is a function, $z_j(t)$, that describes how the shielding of a sample changes through time, which depends on unknown parameters like the rates of geologic processes and the length of time that these processes occurred over. The exposure model is used to calculate nuclide concentrations for a set of the unknown parameters that we compare to the measured abundances. To assess the quality of fit between an exposure model and the measured cosmogenic nuclide concentrations we used an error-weighted, least-squares method. The best-fit values for the unknowns in the exposure model are those that minimize the difference between the modeled and measured concentrations. To determine the uncertainties for these results, we carried out a 10,000 run Monte Carlo simulation that took into account uncertainties in the measured nuclide concentrations.

2.3.2.1. Erosion rates

If we disregard soil mixing, the process of steady exposure and steady degradation since the deposition of the ash would result in a nuclide concentration-depth relationship according to the simple exposure age equation of Lal [1991]. Additionally, we must include any inherited nuclides from before the ash was deposited that might remain in our samples. The concentration of nuclides that we expect is the sum of the inherited nuclides plus those that accumulated after the deposition of the ash:

$$N_{i,j}(N_{i,inh},\varepsilon,t_{ash}) = N_{i,inh} + \frac{P_i(z_j)}{(\lambda_i + \varepsilon/\Lambda)}(1 - e^{-(\lambda_i + \varepsilon/\Lambda)t_{ash}}), \qquad (3)$$

where $N_{i,inh}$ is the concentration of nuclide i inherited from before the ash was deposited that remains in the sediment today, t_{ash} is the depositional age of the ash, and ε is the erosion rate in g cm^{-2} yr^{-1} after the ash is deposited. This argument roughly follows the one in Balco et al. [2005a; 2005b] for till and Putkonen et al. [2008] for regolith. Equation (3) depends on two unknown parameters, $N_{i,inh}$ and ε, and we have at least three measurements of each nuclide at each pit, so we can uniquely determine the unknowns.

Equation (3) assumes that the inherited nuclides from before the ash was deposited are the same at all depths in the pit, and this would only be the case if the colluvium was well mixed before the ash was deposited. Additionally, we can test if the inherited nuclides record erosion before the ash was deposited without any vertical mixing. In this case, the inherited nuclides would fit an exponential profile that decreases exponentially with depth. If we assume that a steadily eroding colluvium with no vertical mixing had reached equilibrium with an erosion rate before the ash was deposited, we can modify equation (3) for an inherited nuclide profile that will fit an exponential profile that has subsequently decayed in the intervening years, and the erosion rate after the ash was deposited, we get:

$$N_{i,j}(\varepsilon_1, \varepsilon_2, t_{ash}) = \frac{P_i(z_j)}{(\lambda_i + \varepsilon_1/\Lambda)} e^{-(\lambda_i) t_{ash}} + \frac{P_i(z_j)}{(\lambda_i + \varepsilon_2/\Lambda)} (1 - e^{-(\lambda_i + \varepsilon_2/\Lambda) t_{ash}}) \quad , \qquad (4)$$

where ε_1 is the erosion rate before the ash is deposited and ε_2 is the erosion rate after the ash is deposited, and the erosion rates have units in g cm^{-2} yr^{-1}. Because the age of the ash is known, we can solve equation (4) for the two unknown erosion rates.

2.3.2.2. Burial history

Hall et al. suggest that the Hart ash may have been buried by ice for a period of time so we need to address the possibility of burial in our exposure model. We can test for the possibility of burial by ice by building an exposure model with three time periods: 1) a steadily eroding surface after the ash is emplaced; 2) a period of burial by ice; 3) a steadily eroding surface after the ice recedes. The total length of time of these three periods is the age of the ash. The concentration of cosmogenic nuclides that we expect in this case is dependent on four unknowns, ε_1 and t_1, the erosion rate and the time period after the ash but before the burial by ice, t_b, the length of time the sediment was buried, and ε_2, the erosion rate after the period of burial. The length of time that the sediment is exposed after any burial by ice, t_2, is set by the age of the ash because the age of the ash must be equal to the three periods of exposure. The concentration of nuclides that we expect from this scenario is:

$$N_{i,j}(\varepsilon_1, \varepsilon_2, t_1, t_b) = \frac{P_i(z_j)}{(\lambda_i + \varepsilon_1/\Lambda)} (1 - e^{-(\lambda_i + \varepsilon_1/\Lambda) t_1}) e^{-(\lambda_i) t_b} e^{-(\lambda_i) t_2} + \frac{P_i(z_j)}{(\lambda_i + \varepsilon_2/\Lambda)} (1 - e^{-(\lambda_i + \varepsilon_2/\Lambda) t_2}) \quad . \qquad (5)$$

Additionally, we can consider inherited nuclides from before the emplacement of the ash in this exposure model following the methods described for equations (3) and (4) above.

2.4. Results

2.4.1. The Arena Valley ash

The cosmogenic nuclide concentrations below the Arena Ash are much lower than expected given the age of the ash. If we were to ignore surface erosion, the concentrations of ^{10}Be and ^{26}Al would yield an exposure age of the Monastery colluvium below the ash of only ~2 Ma, which cannot be true because it is overlain by a 4.33 Ma ash deposit. This indicates that the cosmogenic nuclide concentrations below the ash either reflect the degradation rate of the deposit or burial at this site.

Based on the geologic description of the Arena Valley ash pit, the following geologic events occurred at the sample site. Before the ash was emplaced, a colluvium layer with a well developed desert pavement existed. Because this desert pavement caps the colluvium abruptly, the pavement likely formed as the result of deflation of the underlying colluvium. The desert pavement is not underlain by a layer of eolian sands which would indicate inflation. Thus, the desert pavement that underlies the Arena Valley ash indicates that the surface was undergoing erosion before the ash was deposited. At 4.33 Ma, the Arena Valley ash was deposited on this desert pavement. Today, the Arena Valley ash is capped by a second desert pavement. Though the upper three centimeters of ash contain non-volcanic contaminants, there are no massive deposits of eolian sands which would indicate inflation. Thus, the desert pavement that caps the Arena Valley ash also indicates erosion, though the presence of non-volcanic contaminants in the upper few centimeters of the ash does indicate some infiltration of sand impurities into the massive ash.

From this geologic description, the sample site was experiencing erosion before and after the ash was deposited. Because the age of the ash is known and the two unknowns in this scenario are the two erosion rates before and after the ash was deposited, we begin the analysis by fitting equation (4) to the measured nuclide concentrations. The results from fitting equation (4) to the nuclide measurements are that the erosion rate before the ash was deposited was $2.4 \pm 0.40 \times 10^{-4}$ g cm^{-2} yr^{-1}, and that after the emplacement of the ash the erosion rate was $3.5 \pm 0.41 \times 10^{-5}$ g cm^{-2} yr^{-1}. These rates fit the measurements with a reduced chi square value of 1.6 for all nine data points (Figure 2.4). We can test if the site was experiencing vertical mixing before the ash was deposited by fitting equation (3) to the data, but the fit for this exposure model is not as good as it is for equation (4).

The fact that the ^{10}Be and ^{26}Al measurements fit equation (4) well means that no vertical mixing of the soil has occurred in the period of time that our measurements record. This period is the time required to remove material equivalent to several attenuation lengths of regolith [*Lal*, 1991], which in this case would be over 10 million years. This is too long of a period to look at with ^{10}Be and ^{26}Al because the samples will reach equilibrium between nuclide production and decay in less time than this. The period of time that it takes cosmogenic nuclide concentrations to reach equilibrium with an erosion rate is controlled by the effective half life: $\ln(2)\ (\lambda_i + \varepsilon/\Lambda)^{-1}$, where ε is the erosion rate of the overlying regolith. After a several effective half-lives have passed, the ^{10}Be and ^{26}Al will have reached equilibrium with steady erosion. For an erosion rate of 3.5×10^{-5} g cm^{-2} yr^{-1}, one effective half life of ^{10}Be is ~1 Ma and is ~0.6 Ma for ^{26}Al. The time period that our measurements record is several effective half-lives of the nuclides, which is ~4 Ma for ^{10}Be, or since about the deposition of the Arena Valley ash at 4.33 Ma. Thus, we can say that there has been no vertical mixing of the soil in the past 4.33 Ma at the Arena Valley ash site.

The results from fitting equation (4) to the measurements suggest that the erosion rate before the ash was higher than has been for the last 4.33 Ma. However, the effective half-life of the nuclides suggests that our measurements only record what has happened at this site in the past 4.33 Ma. To further explore the possibility that our samples record information about the erosion rate before the ash was deposited, we begin by assuming that any ^{26}Al from before the ash would have decayed to essentially

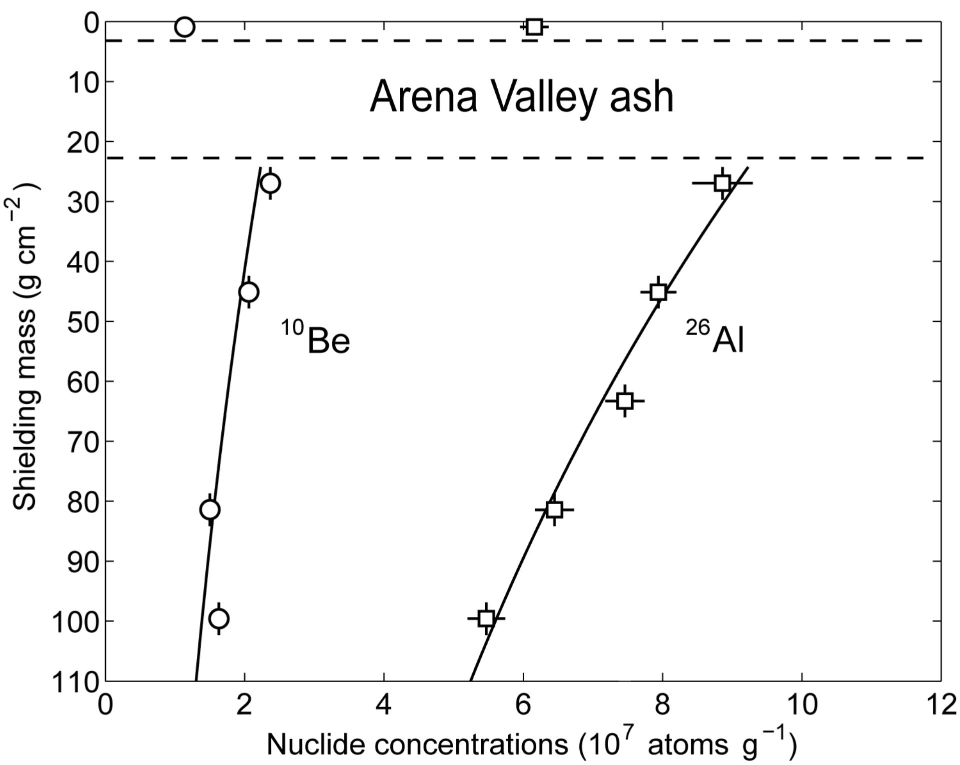

Figure 2.4. Measured ^{10}Be (circles) and ^{26}Al (squares) concentrations (atoms g^{-1}_{quartz}) at the Arena Valley ash site (04-AV-Pit 4). The small vertical and horizontal lines indicate the range of depth of the sediment sampled and the error in the nuclide measurement, respectively. In some cases, these values are smaller than the markers used to indicate each datum. The white box shows the approximate location of the Arena Valley ash in the soil column. The solid lines show the predicted nuclide concentrations for the best-fit model results where the erosion rate before the ash was 2.4 ± 0.40 × 10^{-4} g cm^{-2} yr^{-1}, and that after the emplacement of the ash the erosion rate was 3.5 ± 0.41 × 10^{-5} g cm^{-2} yr^{-1}.

zero in the intervening 4.33 Ma, that the ^{26}Al concentrations in the samples today are the result of only the period of erosion after the ash was deposited. Then, we predict the ^{10}Be concentrations from the erosion rate indicated by the ^{26}Al. Any ^{10}Be in excess of this prediction would then be inherited from before the ash was emplaced.

When we fit an erosion rate to only the ^{26}Al data we get an erosion rate of 4.1×10^{-5} g cm^{-2} yr^{-1} with a reduced chi square fit of 0.71. The ^{10}Be concentrations predicted by this erosion rate fit within the error of our measured concentrations, so there is no information about the erosion rate before the ash was emplaced recorded in these samples, and we cannot confidently say that the erosion rate before the ash was higher than it has been for the past 4.33 Ma. However, if the erosion rate before the ash was lower than it has been since the ash was deposited, then there should be higher ^{10}Be concentrations in our samples than what we measured, so we can conclude that the erosion rate before the ash was not lower than it has been for the past 4.33 Ma.

2.4.2. The Hart ash

The nuclide concentrations of the colluvium below the Hart ash give simple exposure ages of only ~200-500 ka, which is much younger than the overlying 3.9 Ma Hart ash. Thus, the first important result from these cosmogenic nuclide concentrations is that to interpret them as simple exposure ages gives ages that are not in agreement with the known age of the ash and the nuclide concentrations are better interpreted as representing either the degradation rate of the regolith or burial at the site.

The geologic description of the Hart ash site shows that there is a relict sand wedge in the colluvium that underlies the Hart ash. This implies that the colluvium was subject to cryoturbation before the ash was deposited and we would expect some vertical mixing of the sediment as a result of this. The effect of vertical mixing on the cosmogenic nuclide concentrations in the sediment would be to evenly distribute the concentrations to the depth that vertical mixing occurred. The relict sand wedge terminates abruptly at the contact with the Hart ash, which implies that there is an erosional surface at the top of the colluvium and that the site experienced erosion before the ash was deposited. At 3.9 Ma, the Hart ash was deposited on this erosional surface. The Hart ash is now capped by a gravel lag. Though the upper two centimeters of ash contain non-volcanic contaminants, there are no massive deposits of eolian sands which would indicate inflation. Thus, the gravel lag that caps the Hart ash also indicates erosion of the surface, though the presence of non-volcanic sediments in the upper few centimeters of the ash does indicate some infiltration of sediments.

Based on this geologic description, the important geologic events at the Hart ash sample site are the following. Before the Hart ash was deposited, the site was experiencing cryoturbation, which was followed by a period of erosion. The Hart ash was deposited at 3.9 Ma on this erosional surface and has subsequently experienced more erosion after the ash was deposited. The unknowns in this scenario are the erosion rate after the ash was deposited and the inherited nuclide concentrations before the ash was deposited. Because the site was cryoturbated and eroded before the ash, it is unclear if the inherited nuclides from before the Hart ash was deposited will be well mixed or not, so it is appropriate to test if the exposure model described by equation (3) or (4) is a better explanation for the measured nuclide concentrations.

Equation (3) gives a better fit for the measured nuclide concentrations than equation (4), showing that a constant, well-mixed profile for the inherited nuclides fits the measurements better than an exponential profile. The results from equation (3) show that there is essentially zero inherited ^{26}Al, $9.2 \pm 0.67 \times 10^5$ at g^{-1} inherited ^{10}Be remaining in the samples today, and an erosion rate of $4.8 \pm 0.21 \times 10^{-4}$ g cm^{-2} yr^{-1} for the time period that our measurements reflect with a reduced chi square fit of 3.50. The amount of time it would take to remove 3 attenuation lengths of material at this erosion rate is ~1 Ma, and the result is that steady erosion at a rate of 4.8×10^{-4} g cm^{-2} yr^{-1} has gone on for the past 1 Ma. Figure 2.5 shows these results. Because the measured nuclide concentrations fit equation (3) well, no vertical mixing of the soil has occurred in the last 1 Ma.

Given that the Hart ash is 3.9 Ma, it makes sense not to have any inherited ^{26}Al left in these samples. As after almost 4 Ma only ~3% of any inherited ^{26}Al should

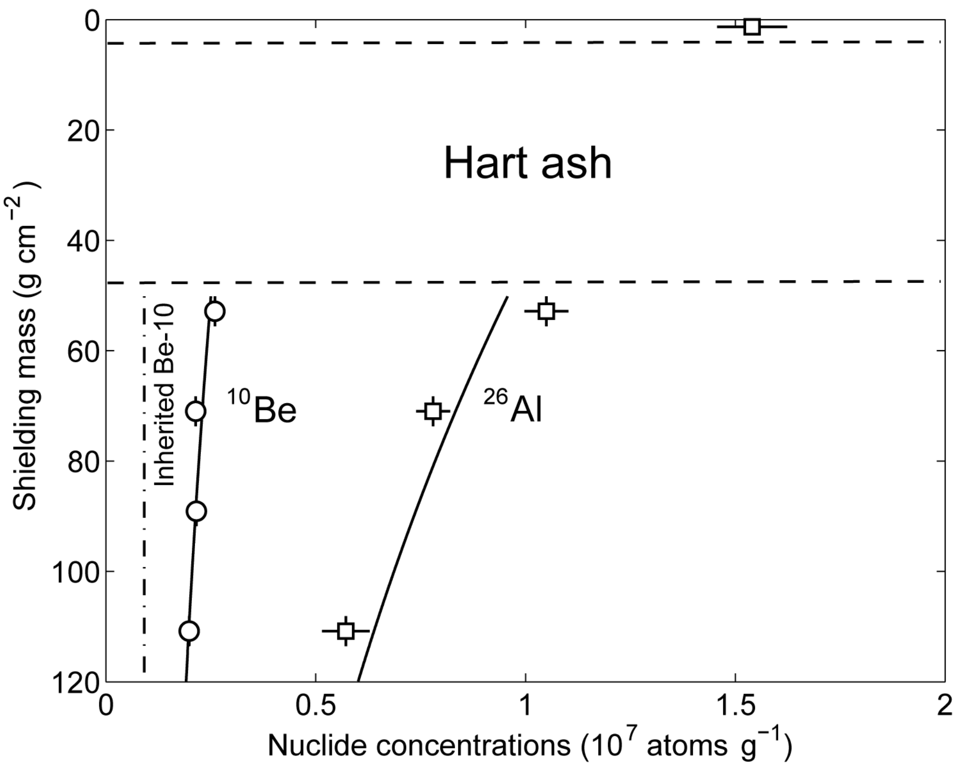

Figure 2.5. Measured ^{10}Be (circles) and ^{26}Al (squares) concentrations (atoms $g^{-1}{}_{quartz}$) at the Hart ash site (04-LWV-Pit 25). The small vertical and horizontal lines indicate the range of depth of the sediment sampled and the error in the nuclide measurement, respectively. In some cases, these values are smaller than the markers used to indicate each datum. The white box shows the approximate location of the Hart ash. The dash-dot vertical line shows the calculated inherited ^{10}Be from before the ash was deposited that remains in the soil today. The solid lines show the best fit model results of zero inherited ^{26}Al, $9.2 \pm 0.67 \times 10^5$ at g^{-1} inherited ^{10}Be remaining in the samples today, and an erosion rate of $4.8 \pm 0.21 \times 10^{-4}$ g cm^{-2} yr^{-1} for the past 3.9 Ma. Because an erosion rate this fast will cause nuclide concentrations to reach a steady state between production and decay in 1 Ma, all that we can conclusively say is that erosion at a rate of $4.8 \pm 0.21 \times 10^{-4}$ g cm^{-2} yr^{-1} has gone on for the past ~1 Ma.

theoretically remain in the samples, and this is smaller than the analytical uncertainties in our ^{26}Al measurements. Following the same reasoning we described for the Arena Valley ash, we can further examine the inherited nuclide concentrations by assuming that the measured ^{26}Al concentrations are the result of erosion for the past 3.9 Ma, and predicting the ^{10}Be concentrations from the erosion rate indicated by the ^{26}Al measurements. The expected erosion rate from the ^{26}Al concentrations is 5.0×10^{-4} g cm^{-2} yr^{-1}, which is slightly higher than the previous fit, but within the error of the results. The ^{10}Be concentrations predicted by this erosion rate are less than the measured ^{10}Be concentrations (Figure 2.6), which is consistent with the idea that there is some ^{10}Be remaining in our samples that is attributable to exposure before the deposition of the ash. If we subtract the predicted ^{10}Be concentrations from the measured ^{10}Be concentrations and assume that the difference is the inherited ^{10}Be, then the resulting inherited ^{10}Be profile is more or less vertical, which is consistent with the idea that the site experienced cryoturbation before the ash was deposited and should be well mixed. This reaffirms the notion that the site was experiencing cryoturbation before the ash was emplaced, resulting in a well-mixed inherited nuclide profile. The nuclide concentrations at the time the ash was deposited predicted by this exercise are higher than the measured concentrations in the samples today, which suggests that the erosion rate before the ash was emplaced at this site was lower than it is today.

Because Hall et al. [1993] suggest that the sample site may have been overridden by a glacier, we fit equation (5) to our data to test for the possibility of burial by ice. In this scenario we have 5 unknowns (inherited Be, two erosion rates, and two time periods) and only 7 data points, so the fit may not be as well constrained as it is in the other exposure models. The results of the best fit scenario are for a well-mixed inherited ^{10}Be concentration of 4.1×10^{6} at g^{-1} at the time the ash was deposited, an erosion rate of 3.1×10^{-4} g cm^{-2} yr^{-1} for the 0.44 Ma after the ash was deposited, a burial period of 2.2 Ma, followed by an erosion rate of 5.0×10^{-4} g cm^{-2} yr^{-1} for 1.3 Ma after the burial until today. These results fit the data with a reduced chi square value of 6.72. Though the chi square fit is not as good for this exposure model, the result that

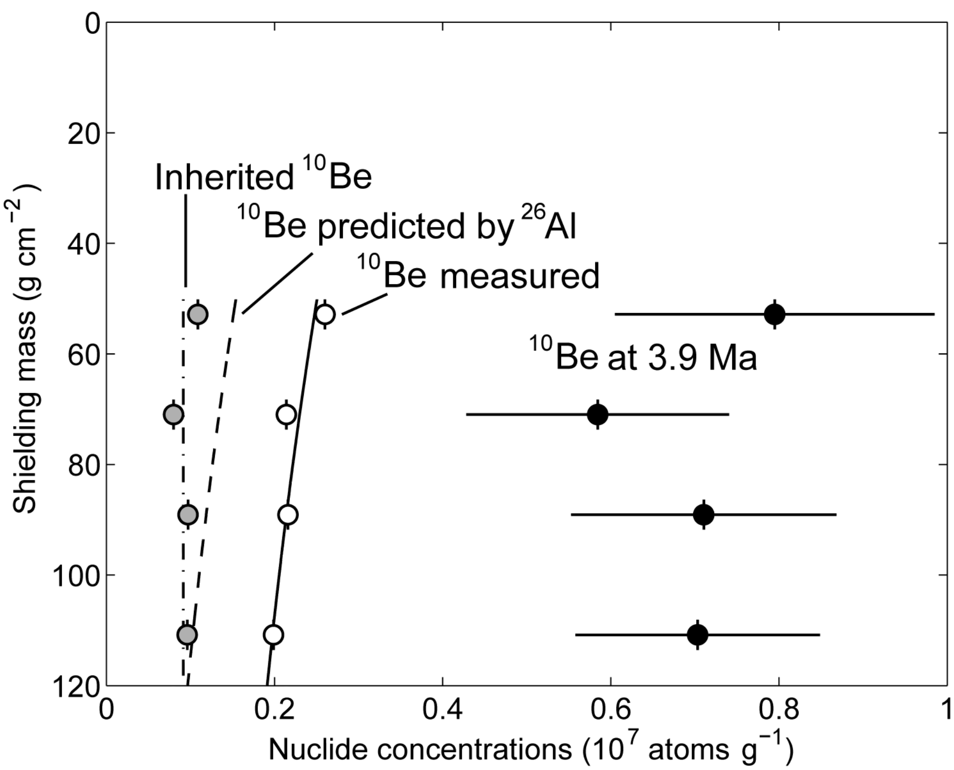

Figure 2.6. Analyzing the inherited ^{10}Be concentrations below the Hart ash. The measured concentrations of ^{10}Be are shown by the open circles. The solid line shows the best fit line through the ^{10}Be data and the vertical dash-dot line shows the predicted well-mixed inherited ^{10}Be remaining in the sediment today (9.2×10^5 atoms g^{-1}) from the analysis shown in Figure 2.5. The dashed line shows the amount of ^{10}Be predicted by the best fit erosion rate for the ^{26}Al concentrations. The gray circles are the difference between the measured ^{10}Be concentration and the ^{10}Be concentration predicted by the ^{26}Al erosion rate, which in this analysis is assumed to be the inherited ^{10}Be left in these samples. These calculated inherited ^{10}Be concentrations nearly fall along the well-mixed inheritance line. The black circles show the value of the inherited ^{10}Be at 3.9 Ma when the Hart ash was deposited. That is, if we decay the ^{10}Be concentrations in the black circles for 3.9 Ma, we get the values of the grey circles. The nuclide concentrations at the time the ash was deposited predicted by this exercise are higher than the measured concentrations in the samples today, which may indicate that erosion at this site before the ash was emplaced was lower than it is today.

there has been erosion of ~5×10^{-4} g cm^{-2} yr^{-1} is consistent with the results from the other exposure model. ^{10}Be and ^{26}Al concentrations will come into equilibrium with an erosion rate of ~5×10^{-4} g cm^{-2} yr^{-1} in ~1 Ma, effectively removing any information that we have about burial, so we can only conclusively say that erosion at this rate has gone on for at the last ~1 Ma.

2.4.3. Samples above the ash

The final observation is that the nuclide concentrations for the samples above the ash at both sites are different than what is predicted by the erosion rate indicated by the samples below the ash. Stratigraphically, these samples are different from the ash layers that they overlie, so their exposure history is not necessarily related to either the underlying ash or colluvium. At both sample sites, the samples above the ash came from the uppermost layer of ash that was mixed with quartz sands. Our interpretation of the surface samples is that they represent material that is actively moving at the surface today, as observed by Putkonen at al. [2007]. This material could be moving downslope or it could be wind-blown, but regardless of the method of transportation, this upper sample contains cosmogenic nuclide concentrations that are unrelated to those that are below the ash.

At the Arena Valley ash site, the simple exposure age for the sample from 0-1 cm depth is ~600 ka, which corresponds to an erosion rate of 1.8×10^{-4} g cm^{-2} yr^{-1}., which is much higher than the erosion rate indicated by the samples below the ash. Because the erosion rate of the surface sample is so high, we can not create an exposure model where the samples below the ash accumulate enough nuclides in the first 1.8 Ma of exposure to reach their measured concentrations. The result is that the erosion rate of the surface sample can not have been the erosion rate of the entire sediment package for the past 2.6 Ma, which would be the period of time that this surface sample records. Thus, the nuclide concentrations of the sample above the ash are unrelated to those from the colluvium below the ash.

For the sample above the Hart ash, we only have a ^{26}Al measurement and it is a higher concentration than expected given the steady erosion rate determined from the colluvium below the ash. The ^{26}Al concentration of this sample gives a simple exposure age of 280 ka and an erosion rate of 4.2×10^{-4} g cm^{-2} yr^{-1}. This erosion rate is ~20% lower than the best fit erosion rate for the sample below the ash. If this is the erosion rate that has gone on for the past 1.15 Ma (the time period that this sample records), then the concentration of ^{26}Al in the samples below the ash would have to be much higher than measured. Again, our interpretation of the surface sample is that it represents material that is actively being transported, and is unrelated stratigraphically to the material below it.

2.5. Discussion

2.5.1. Erosion rates of regolith in the MDV and implications for landscape evolution

At both of the ash sites, the concentration of ^{10}Be and ^{26}Al both above and below the ash deposits are lower than expected given the ages of the ashes, burial is not indicated by the exposure models or the ^{26}Al/^{10}Be ratio, and the nuclide concentrations are best interpreted as degradation rates. If we know the density of the eroding regolith we can convert these erosion rates into a surface lowering rate. Assuming that the density of the material removed was at least that of the ash and no greater than the colluvium, we can put a range on the amount of material removed at each site. Using an average density of 1.01 g cm^{-3} for the ash and 1.81 g cm^{-3} for the regolith (Table 1) we calculate an erosion rate of 0.19 – 0.35 m Ma^{-1} at the Arena Valley ash site and 2.6 – 4.7 m Ma^{-1} at the Hart ash site. Because we only know that erosion has been occurring at the Arena Valley ash site for 4.33 Ma and at the Hart ash site for 1 Ma, the net removal of material over these time periods is 0.84 – 1.5 m at the Arena Valley ash site, and 2.6-4.7 m at the Hart ash site. The erosion rates determined in this study are similar to those found by Putkonen et al. [2008] who found that a landslide deposit in Arena Valley was eroding at a rate of 2.1 m Ma^{-1} for at least the past ~2 Ma. At the Hart ash site, our results are consistent with those found by Schiller et al. [2009], who used atmospherically produced ^{10}Be to determine erosion rates of only 0.5 m Ma^{-1} before the Hart ash was deposited, and 2.8 m Ma^{-1} after the ash was deposited.

There is an order of magnitude difference in the erosion rates between the Arena Valley and Hart ash sites, and elevation, distance from the coast, mean annual temperature, relative humidity, exposure to and protection from high velocity winds all affect erosion rates in the MDV [*Marchant and Head*, 2007]. Distinguishing the relative effects of these processes at each site is beyond the scope of this paper, but our results are consistent with the notion that microclimate variations in the MDV will result in different erosion rates [*e.g. Marchant and Denton*, 1996; *Marchant and Head*, 2007] because the erosion rates in the stable upland zone (where the Arena Valley ash lies) are lower than those in the inland mixed zone (where the Hart ash lies). Additionally, our results suggest that the higher erosion rate at the Hart ash site are due to the fact that the site is capped by a gravel lag consisting of granitoid pebbles that weathers more quickly and provides less armor than the dolerite desert pavement that caps the Arena Valley ash. This notion is consistent with the idea proposed by Lancaster [2002] that erosion rates in the MDV are more influenced by the availability of material than by wind speed.

The presence of millions of years old ash deposits at or near the surface in the MDV is often geomorphologically interpreted as an indication that little to no slope evolution has occurred since the ash was deposited and that the surface in the stable upland zone "is essentially paralyzed under persistent hyper-arid, cold-desert conditions" [*Marchant and Denton*, 1996]. The degradation rates determined in this study are small when compared to erosion rates on a global scale [*e.g. Bierman and Nichols*, 2004], but even these slow erosion rates can lower the regolith surfaces in the order of meters over the long life spans of these surfaces. Creep is a common process of soil erosion in a variety of environments, and in temperate regions has been shown to move soil downslope at rates of ~100 m ka^{-1} [*Heimsath et al.*, 2002]. This study reaffirms the notion that the MDV are a geomorphologically unique site where degradation is slow and does not involve creep. However, ashes at the surface are not a clear indication that there has never been more material above the ash [*Bockheim and Ackert*, 2007]. Our results show that there has been erosion of ash or regolith above the ash, and that the material that now caps the ash deposits lies disconformably on the ash.

2.5.2. Implications for glacial history

Based on the observations of Hall et al. [1993] it is possible that the Hart ash has been covered by ice at some point after it was deposited. Though the ash appears to be a primary airfall deposit and it seems that any glacial overriding would disturb an ash deposit, there is precedent in the MDV for glaciers to cover an ash and leave it undisturbed. The Mt. Thundergut ash (DMS90-124B) in Nibelungen Valley in the Asgard Range, first described by Marchant et al. [1993b] and in more detail by Marchant et al. [1996] is a 2 cm thick ash layer that overlies a desert pavement. The ash is capped by 60 cm of till and appears to be unweathered because it contains less than 5% clay-sized grains. This ash is interpreted as a primary airfall ash that lies in situ because it lacks nonvolcanic contaminants and preserves glass shards. Based on the observations by Hall et al. and the precedent set by the Mt. Thundergut ash, it is plausible that weakly erosive cold-based ice once covered the Hart ash site and did not disturb the basal layers of the Hart ash.

The best fit exposure model for a burial scenario at the Hart ash site results in the site being buried for 2.2 Ma, followed by erosion for 1.3 Ma after burial until today at a rate of 5.0×10^{-4} g cm^{-2} yr^{-1}. This erosion rate is within the error of the erosion rate for the exposure model with erosion and no burial and would remove three attenuation lengths of material in ~1 Ma, so the dominant and persistent signature in the nuclide concentrations is erosion for past ~1 Ma. Based on the surrounding geomorphology, the Hart ash site does not appear to have been covered by ice or water since it was deposited. Hall et al. [1993] extensively mapped the glacial deposits from the local alpine glaciers, and the Hart ash site lies outside of the limit that these glaciers reached. The Hart ash site is also above the 300 m limit of fjord deposits found in Wright Valley [*B.L. Hall et al.*, 1993; *Prentice and Krusic*, 2005]. Prentice and Krusic [2005] limit the last expansion of ice in Lower Wright Valley to between 5.5 ± 0.4 - 3.7 ± 0.1 Ma. Thus, though the scenario that the site has been covered by cold based ice is mathematically possible, it is not necessary and all that we can conclusively say is that the Hart ash site has been eroding at a rate of $4.8 \pm 0.21 \times 10^{-4}$ g cm^{-2} yr^{-1} for the past 1 Ma.

The climate interpretation based on the ash deposits found in the MDV is that the cold and dry polar conditions in the MDV today have persisted since the oldest of the ashes was deposited, ~ 15 Ma. This interpretation is based mainly on the unweathered state of the ashes and the argument is that if the climate was ever warmer or wetter, these ashes would show more weathering. However, erosion at these sites may have removed the most weathered parts of the ash. If the upper layers of the ash were weathered, erosion would have removed these layers, and the record of weathering would have been removed from the site. Similarly, if the Hart ash site had been buried by ice and a thin till layer was deposited above the ash, erosion may have removed the till in the intervening 1 Ma ice free period leaving no surface evidence that the site was ever covered by ice. Though we do not suggest that any of these events necessarily did occur at either of the ash sites, we want to emphasize that erosion has removed between 0.84 – 4.7 m of the geologic record at these sites, which means that we have lost part of the geologic record of what happened after the ashes were deposited.

2.6. Conclusions

We examined how regolith has degraded in a cold polar desert for the past 1 – 4.33 Ma. At both of the ash sites, the concentration of ^{10}Be and ^{26}Al both above and below the ash deposits are lower than expected given the ages of the ashes. Burial by weakly-erosive, cold-based ice cannot explain these low nuclide concentrations at the Arena Valley ash site, and at the Hart ash site, burial is possible, but is indistinguishable from erosion for the past 1 Ma. Therefore, the nuclide concentrations are consistent with the depositional ages of the ashes only if they are interpreted as the result of degradation at these sites. The best-fit erosion rates for the two sites are 3.5 ± 0.41 × 10^{-5} g cm^{-2} yr^{-1} (0.19 m Ma^{-1} of regolith or 0.35 m Ma^{-1} of ash) for the past 4.33 Ma at the Arena Valley ash site, and an erosion rate of 4.8 ± 0.21 × 10^{-4} g cm^{-2} yr^{-1} (2.6 of regolith or 4.7 m Ma^{-1} of ash) for the past 1 Ma at the Hart ash site. While these rates are slow, they are measurable, and given an additional 0.1 – 0.5 Ma the ash deposits would be entirely removed from these sites. At both of these sites there is no indication of soil creep or vertical mixing of the sediment in the last 1 – 4.33 Ma. The cosmogenic nuclide concentrations in the samples above the ash do not have the same exposure history as those below the ash, supporting the idea that regolith degradation in the MDV occurs without creep and is limited to the upper few centimeters of the regolith. Collectively these results paint a picture of a more dynamic environment than previously thought and suggest that sizeable fraction of the sedimentary record may have been removed from the landscape.

Chapter 3

DEGRADATION OF GLACIAL DEPOSITS QUANTIFIED WITH COSMOGENIC NUCLIDES, QUARTERMAIN MOUNTAINS, ANTARCTICA

3.1. Introduction

The Quartermain Mountains are located in the southwestern section of the McMurdo Dry Valleys (MDV) region of Antarctica (Figure 3.1), and they contain some of the oldest glacial deposits in Antarctica [*e.g. Denton et al.*, 1984, 1993]. The glacial stratigraphy recorded in the Quartermain Mountains is central to our understanding of the history of the East Antarctic Ice Sheet for the past 11.3 Ma [*Marchant et al.*, 1993a]. Chronologies for glaciogenic deposits in the Quartermain Mountains have been established with cosmogenic nuclides [*e.g. Brown et al.*, 1991; *Brook et al.*, 1993, 1995; *Schafer et al.*, 1999] and with $^{40}Ar/^{39}Ar$ ages of *in-situ* volcanic ashes [*e.g. Marchant et al.*, 1993a; *Sugden et al.*, 1995]. Many of the glacial deposits are millions of years old, yet they have maintained their meter-scale morphology despite the fact that bedrock and regolith erosion rates in the Quartermain Mountains are 0.1 – 4m Ma^{-1} [*Summerfield et al.*, 1999; *Putkonen et al.*, 2008a].

An intriguing feature of glacial deposits in the Quartermain Mountains is that buried glacial ice underlies the Granite Drift in central Beacon Valley [*Sugden et al.*, 1995]. Dates for the Granite Drift range from 310 ka [*Ng et al.*, 2005] to 8.1 Ma [*Sugden et al.*, 1995; *Marchant et al.*, 2002]. In contrast, the Taylor and Quartermain drifts in Arena valley lie on weathered sandstone bedrock and mostly lack sufficient pore moisture to form ice cement in the soil. There is considerable ongoing debate regarding the age and stability of ground ice in the MDV. Presently, sublimation rates of bare ice in the MDV are on the order of 10^4 – 10^5 m Ma^{-1} [*Clow et al.*, 1988]. Modeled and measured sublimation rates of ice in soils suggest that without any recharge mechanisms ground ice should sublime in the upper few meters of soil on the order of 10^3-10^5 years, yet ground ice persists near the surface within Miocene-aged

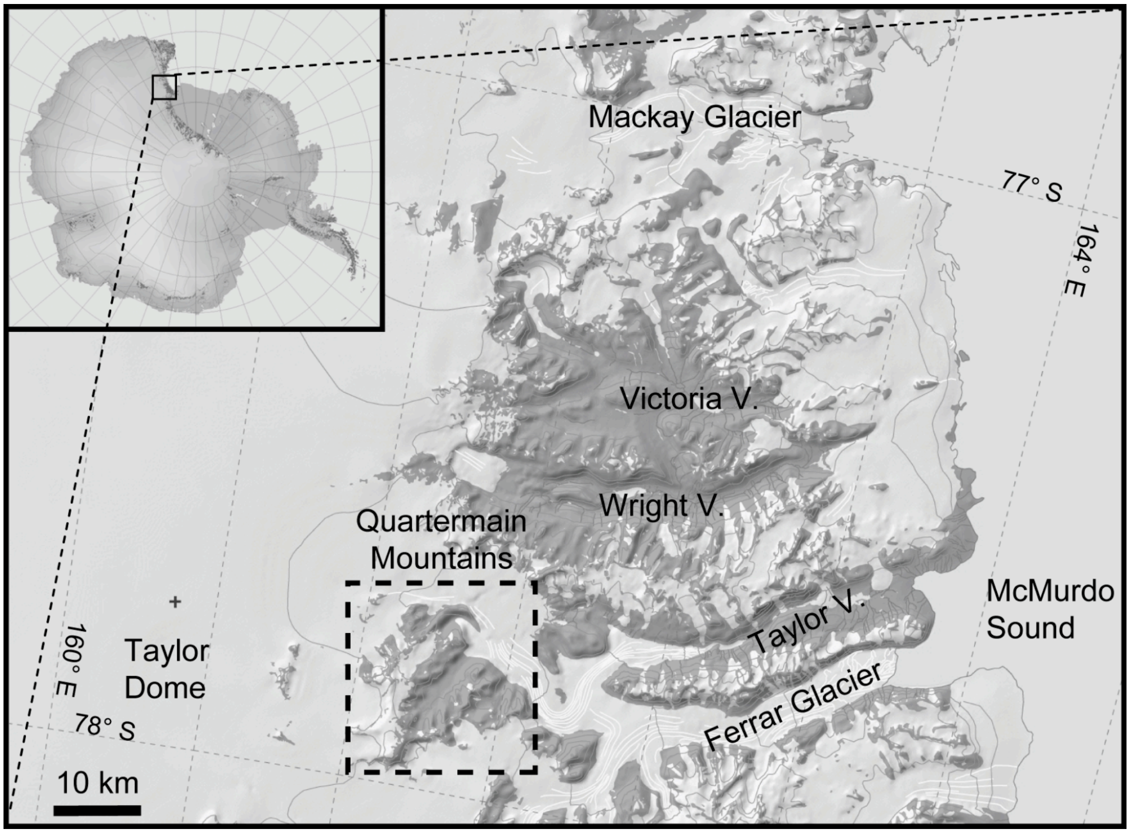

Figure 3.1. Map of the McMurdo Dry Valleys, Antarctica. The Quartermain Mountains, the subject of this study, are shown in the dashed box.

soils in the MDV [*McKay et al.*, 1998; *Hindmarsh et al.*, 1998; *Hagedorn et al.*, 2007; *Campbell and Claridge*, 2006; *Bockheim et al.*, 2007].

The primary objective of this paper is to quantify degradation rates for glacial deposits in the Quartermain Mountains. In this paper, we will use the term degradation to mean the general lowering of the earth's surface, regardless of process. We are motivated by the apparent conflict between the persistence of landforms that are eroding at rates on the order of a few meters per million years while maintaining their meter-scale morphology, and the presence of ground ice in Miocene-aged soils in the MDV. To accomplish our objective, we will use the concentration of cosmogenic nuclides ^{10}Be and ^{26}Al from meter deep vertical sections in glacial tills to examine how quickly and by what geomorphic processes glacial deposits in the Quartermain Mountains degrade.

3.2. Field area

The Quartermain Mountains, Antarctica contain five main valleys, the largest of which are Beacon (215 km^2) and Arena (68 km^2) (Figure 3.2). Bedrock in the valleys consists of Devonian-Triassic sandstones of the Beacon Supergroup and Jurassic sills and dikes of the Ferrar Dolerite [*Barrett*, 1981]. The general climate of the area is a cold, hyperarid, polar desert. The mean annual temperature on the floor of Beacon Valley is -24°C [*Putkonen et al.*, 2003], and water-equivalent precipitation is probably less than 10 mm yr^{-1} [*Bockheim*, 2007]. At Beacon Valley, mean daily wind speed over a three-year period was 4.2 m s^{-1}, the mean of maximum daily 15 minute wind speeds was 7.6 m s^{-1}, and the maximum 15 minute wind speed was 22 m s^{-1} [J. Putkonen, personal communication, 2009].

The Quartermain Mountains contain a number of well mapped glaciogenic surface deposits and bedrock landforms that reflect the expansion of Taylor Glacier, overriding of the entire range by NE-flowing ice, and the growth of local alpine glaciers [*Denton et al.*, 1984; *Marchant et al.*, 1993a]. The mouth of Arena Valley shows a series of arcuate moraines that open to Taylor Glacier and boulders on these

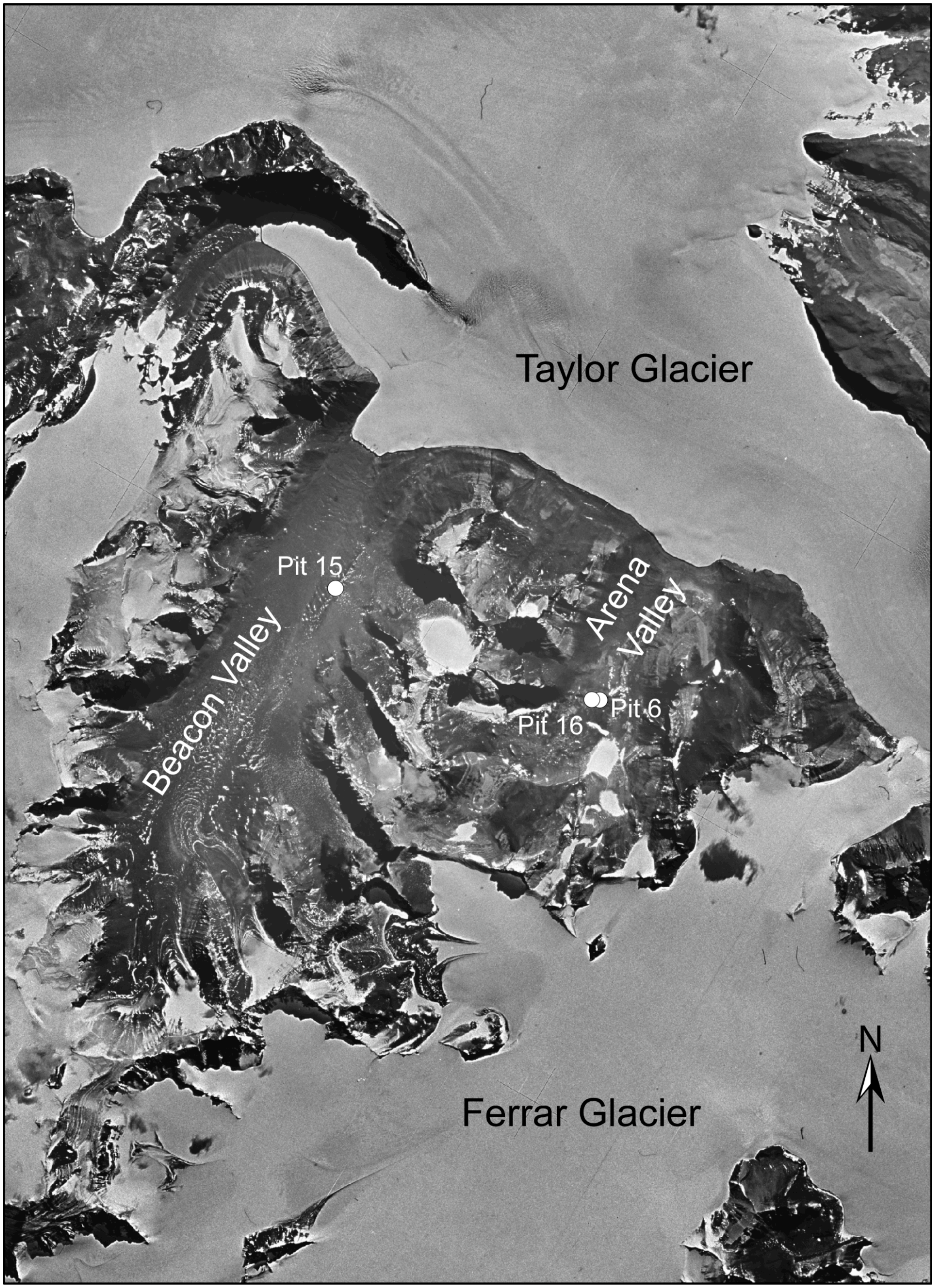

Figure 3.2. Satellite image of the Quartermain Mountains (scale 1:100,000) showing the three sample sites.

moraines have been dated with cosmogenic nuclides and yield ages from 0.12 – 2.2 Ma [*Brown et al.*, 1991; *Brook et al.*, 1993]. Boulders from the Quartermain I till that reflect older expansions of the Taylor glacier have been dated up to 3.5 Ma by Brook et al. [1995]. With detailed geomorphic mapping and $^{40}Ar/^{39}Ar$ ages of *in-situ* ash deposits, Marchant et al. [1993a] determined relative and absolute chronologies for most of the surface deposits in Arena Valley and found that the oldest deposits in Arena valley are > 11.3 Ma. In Beacon Valley, Sugden et al. [1995] found relict glacial ice overlain by a 50 cm thick deposit of sublimation till that itself contains 8.1 Ma reworked ash deposits in relict sand wedges. Clasts above this glacial ice and on local debris-covered glaciers have been dated with cosmogenic isotopes and the oldest of these samples yield minimum ages of 1.18 Ma [*Marchant et al.*, 2002] and 2.3 Ma [*Schäfer et al.*, 2000], suggesting that debris-covered ice can persist for millions of years. The glacial stratigraphy recorded in the Quartermain Mountains is central to the argument that the MDV have remained cold and hyperarid since the mid-Miocene [*e.g. Sugden et al.*, 1993]

3.2.1. Sample sites

We sampled three different glacial deposits at three separate sites in the Quartermain Mountains (Figure 3.2). In Beacon Valley, we sampled an undated moraine on the southwest flank of the mouth of the valley (04-BV-Pit 15: 77.82970°S, 160.69083°E, 1370m). This moraine likely corresponds with one of the series of Taylor moraines in Arena Valley, and based on the soil properties of this till, Bockheim [2007] correlated this moraine with either the Taylor IVa (1.0 Ma) or IVb (2.2 Ma) moraines. The sample site is at the crest of the moraine, and we collected four samples beginning from the bottom of the desert pavement that armors the surface of the moraine to a depth of 60 cm.

In Arena Valley we sampled the Arena and Quartermain I tills. These tills have been described in detail by Marchant et al. [1993a], who found that the Quartermain I till overlies the Arena till with a sharp stratigraphic contact. Brown et al. [1991]

initially dated the Quartermain I till as > 4.4 Ma, but Brook et al. [1995] later revised the age of this till to > 3.5 Ma. Based on stratigraphic relationships and soil properties, Marchant et al. [1993a] found that the Arena till is > 11.3 Ma. We sampled both tills at a site that showed this stratigraphic contact (04-AV-Pit 6: 77.85500°S, 160.95928°E, 1350m). At this site, a buried *in-situ* desert pavement caps the underlying Arena till. We collected three samples in a vertical section from each till at this site for a total of six samples. At another site (04-AV-Pit 16: 77.85487°S, 160.95110°E, 1329 m), the Arena till crops out at the surface farther down Arena Valley, and we collected four samples from a vertical section of the Arena till at this site.

3.3. Methods

3.3.1. Sample collection and processing

In the field, we collected a series of bulk sediment samples from a hand-dug soil pit. At all sample sites we collected a sediment sample from just below the modern desert pavement that armored each site, and a series of till samples at various depths below the ground surface. We did not encounter ice-cemented soil at any of the sites. Site elevations were determined by repeated barometric pressure measurements, which were tied to differential GPS markers. The cosmogenic exposure geometry of each site was determined with a clinometer and compass.

Before isolating quartz from the samples for cosmogenic nuclide analysis, we measured the density of the tills by packing the sediment samples into a container of known volume. For quartz extraction, we used the 0.3-0.5 mm grain size fraction, which was purified by heavy-liquid mineral separation and repeated etching in 2% HF. We extracted beryllium and aluminum from the quartz using standard methods [*Stone*, 2004] and measured ^{10}Be/^{9}Be and ^{27}Al/^{26}Al isotope ratios at the PRIME lab at Purdue University in West Lafayette, IN. The combined carrier and process blanks contained $1.67 \pm 0.462 \times 10^5$ atoms ^{10}Be and $0 \pm 2.04 \times 10^5$ atoms ^{26}Al. These are always less than 0.4% of the measured ^{10}Be atoms and 0.1% of the measured ^{26}Al atoms. The beryllium measurements were originally standardized to the ICN standard, and we have re-standardized them to the 07KNSTD [*Nishiizumi et al.*, 2007]. For aluminum, the isotope ratios were referenced to the Nishiizumi [2004] standard. The sample depths, densities, and nuclide concentrations are found in Table 1. We determined the ^{10}Be and ^{26}Al production rates at the sample sites using the method of Stone [2000]. ^{10}Be production rates were further reduced by a factor of 1.106 following Nishiizumi et al. [2007]. Nuclide production rates for each site are given in Table 2.

3.3.2. Cosmogenic nuclide concentration analysis

The general concept for analyzing the cosmogenic nuclide measurements is to use the local geologic and geomorphic context to determine what happened geologically. Then we construct an exposure model to describe these geologic events and use the ^{10}Be and ^{26}Al concentrations to determine when and how fast the geologic events occurred. The concentration of cosmogenic nuclides, $N_{i,j}$, in samples that are experiencing steady erosion can be described by [*Lal*, 1991]:

$$N_{i,j}(\varepsilon,t) = N_{i,inh} + \frac{P_i \cdot e^{\frac{-z_j}{\Lambda}}}{(\lambda_i + \varepsilon/\Lambda)}(1 - e^{-(\lambda_i + \varepsilon/\Lambda)t}) \quad . \qquad (1)$$

The subscript *i* refer to the nuclide of interest, either ^{10}Be or ^{26}Al, and the subscript *j* refer to the individual samples at each site. *N* is the concentration of the nuclide (atoms g^{-1}), $N_{i,inh}$ is the inherited concentration of the nuclides that remain in the till today (atoms g^{-1}), P_i is the surface production rate of nuclide *i* (atoms g^{-1} yr^{-1}), *z* is the shielding mass of sample *j* (g cm^{-2}), λ is the decay constant for nuclide *i* (yr^{-1}), ε is the erosion rate (g cm^{-2} yr^{-1}), Λ is the attenuation length (150 g cm^{-2}), and *t* is the exposure duration (yr). In this study, we have used decay constants of $\lambda_{10} = 5.1 \times 10^{-7}$ yr^{-1} for ^{10}Be [*Nishiizumi et al.*, 2007] and $\lambda_{26} = 9.78 \times 10^{-7}$ yr^{-1} for ^{26}Al [*Nishiizumi*, 2004]. Because the depth of each sample is less than one meter, we will assume that production of ^{10}Be and ^{26}Al is entirely by spallation [*Gosse and Phillips*, 2001].

Table 3.1. Depth, density, effective shielding mass, and isotope data for the samples.

Sample ID	Depth in soil (cm)	Soil density ($g \cdot cm^{-3}$)	Effective shielding mass ($g \cdot cm^{-2}$)	$^{10}Be \cdot 10^6 \pm 1$ std uncertainty ($at \cdot g^{-1}_{quartz}$)	$^{26}Al \cdot 10^6 \pm 1$ std uncertainty ($at \cdot g^{-1}_{quartz}$)
04-BV-Pit 15: Beacon moraine					
04-BV-Pit 15 0-3	0-3	1.78	2.75	11.8 ± 0.220	61.7 ± 2.18
04-BV-Pit 15 18-22	18-22	1.90	36.8	9.24 ± 0.228	49.1 ± 1.87
04-BV-Pit 15 35-40	35-40	1.89	69.0	7.02 ± 0.202	38.0 ± 1.38
04-BV-Pit 15 55-60	55-60	1.81	106	5.55 ± 0.148	25.3 ± 1.20
	Mean =	1.84			
04-AV-Pit 6: Quartermain till overlies Arena till					
04-AV-Pit 6 2-5	2-5	1.90	6.6	19.8 ± 0.895	81.6 ± 2.58
04-AV-Pit 6 18-22	18-22	1.98	37.6	16.2 ± 0.478	66.5 ± 2.51
04-AV-Pit 6 30-35	30-35	2.07	61.1	12.3 ± 0.320	51.9 ± 1.83
04-AV-Pit 6 33-37	33-37	1.74	65.8	10.4 ± 0.289	47.2 ± 1.79
04-AV-Pit 6 66-69	66-69	1.78	127	7.24 ± 0.295	32.7 ± 1.51
04-AV-Pit 6 84-88	84-88	1.80	162	5.95 ± 0.197	26.4 ± 1.04
	Mean =	1.88			
04-AV-Pit 16: Arena till					
04-AV-Pit 16 0-2	0-2	1.70	1.69	10.0 ± 0.328	58.1 ± 1.65
04-AV-Pit 16 14-16	14-16	1.66	25.4	9.52 ± 0.263	45.7 ± 2.09
04-AV-Pit 16 29-31	29-31	1.71	50.7	6.78 ± 0.205	36.7 ± 1.57
04-AV-Pit 16 52-54	52-54	1.70	89.6	5.58 ± 0.139	29.8 ± 1.11
	Mean =	1.69			

Table 3.2. Nuclide production values at each sample site.

Sample Site	Latitude (°S)	Longitude (°E)	Elevaion (masl)	Shielding correction	P_{10} ($at \cdot g^{-1} \cdot yr^{-1}$)	P_{26} ($at \cdot g^{-1} \cdot yr^{-1}$)
04-AV-Pit 6	77.85500	160.95928	1350	0.994	19.4	131.5
04-BV-Pit 15	77.82970	160.69083	1370	0.996	19.7	133.8
04-AV-Pit 16	77.85487	160.95110	1329	0.993	19.0	129.1

Because many moraines in the MDV are initially deposited as ice-cored deposits [*Debenham*, 1921; *Hall et al.*, 1993; *Denton and Marchant*, 2000] and because the Granite Drift in Beacon Valley is a sublimation till that formed as a lag as the underlying ice sublimed away [*Sugden et al.*, 1995], we must consider the possibility that at sometime in the past the glacial deposits we sampled contained either massive ice or ground ice that has subsequently sublimed away. Exposure models considering the buildup of till by the sublimation of dirty ice have been described by Schafer et al. [2000] and Ng et al. [2005], but neither of these studies considered the case that the till was also undergoing erosion. Because all three of our sample sites are abruptly capped by a desert pavement, indicating that the pavement formed as a lag during erosion at the site, it is apparent that these sites have experienced erosion, so we must include erosion in the model that describes the sublimation of debris-laden ice.

Figure 3.3 shows how the shielding mass above a given sample changes through time for the sublimation of dirty ice and the erosion of the resulting till. The samples that we collected today were initially deposited at a greater depth and were shielded by ice that has sublimed away and by more debris that has eroded away. While the samples were in ice, they approached the surface at a rate equal to the ice sublimation rate plus the surface erosion rate because both of these processes were occurring simultaneously. Once the ice front moves past a sample, the sample accretes into the till, and the sample approaches the surface at a rate equal to only the surface erosion rate.

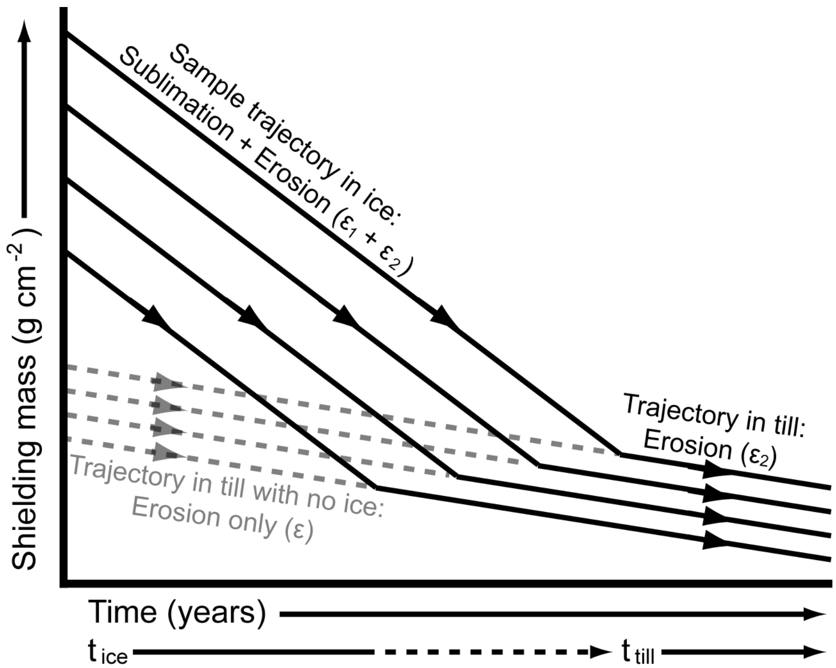

Figure 3.3. The shielding vs. time plot for the sublimation exposure model in a reference frame that is fixed to the surface. The solid lines denote the mass that shields each sample through time. Initially, the samples are deposited in debris-rich ice, so they are shielded by both the overlying debris and ice. After this unit is deposited, it degrades due to both sublimation of the ice and erosion of the resulting till, and the samples approach the surface at a rate equal to the sum of these processes. Samples remain in the ice until the ice front has passed below the sample, at which point the sample accretes into the till. Once a sample has accreted into the till, the only process that affects the shielding of the sample is erosion, and the samples will approach the surface at a rate equal to erosion alone. The dashed lines that connect the erosion lines to the y-axis indicate the initial amount of debris that would have shielded each sample when the unit was first deposited.

The concentration of nuclides predicted by this exposure model are the sum of the inherited nuclides remaining in the till today, plus the nuclides accumulated while the sample was in ice (which will begin to decay once the sample is accreted into the till), plus the nuclides accumulated while the sample was in the till. This can be described by the following equation:

$$N_{i,j}(\varepsilon_1,\varepsilon_2,t_{ice},t_{till}) = N_{i,inh} + \frac{P_i * e^{\frac{-z_j}{\Lambda}}}{(\lambda_i + {}^{(\varepsilon_1+\varepsilon_2)}\!/\!_{\Lambda})}(1 - e^{-(\lambda_i + {}^{(\varepsilon_1+\varepsilon_2)}\!/\!_{\Lambda})t_{ice}})e^{-\lambda_i \cdot t_{till}} + \ldots$$

$$\frac{P_i * e^{\frac{-z_j}{\Lambda}}}{(\lambda_i + {}^{\varepsilon_2}\!/\!_{\Lambda})}(1 - e^{-(\lambda_i + {}^{\varepsilon_2}\!/\!_{\Lambda})t_{till}})\,. \qquad (2)$$

Here ε_1 is the sublimation rate and ε_2 is the erosion rate (g cm^{-2} yr^{-1}). The two time variables, t_{ice} and t_{till}, are the amount of time that each sample spends in the ice and till, respectively, and must sum to equal the age of the deposit t_f. To solve equation (2), we must know t_{ice} and t_{till} for each sample, and to determine these values we need to know the original condition of the deposit, including the amount of ice and the concentration of debris in this ice, C_0.

We reconstruct the original condition of the deposit by finding the combination of sublimation rate, erosion rate, concentration of debris in the ice, and age of the deposit that results in the samples having the shielding mass where we collected them. Once we know the initial amount of ice that shielded each sample, then, the amount of time that each sample spends in ice, t_{ice}, is the amount of time it takes to remove this ice for some sublimation rate, ε_1. Thus, t_{ice}, can be described by:

$$t_{ice}(\varepsilon_1,\varepsilon_2,C_0,t_f) = \frac{(z_j + \varepsilon_2 \cdot t_f)\cdot(1 - C_0)\cdot\rho_{ice}}{C_0 \cdot \rho_{rock} \cdot \varepsilon_1}\,. \qquad (3)$$

In equation (3), z_j is the shielding mass of sample i where we collected it, ρ_{ice} is 0.92 g cm^{-3} and ρ_{rock} is 2.85 g cm^{-3}, which assumes that the till is a mixture of quartz (ρ = 2.65

g cm^{-3}) and dolerite (ρ = 3 g cm^{-3}). In this case, the density of the rock material in the deposit is used because the deposit is initially an ice-supported matrix. The time that each sample spends in the till is simply $t_{till} = t_f - t_{ice}$. Thus, to solve equation (2) for each nuclide we really have five unknowns: the inherited ^{10}Be or ^{26}Al, the sublimation rate, the erosion rate, the initial debris concentration in the ice and the age of the deposit.

There are some additional constraints that need to be added to this exposure model to make it physically valid. One constraint on the model is that the sublimation rate needs to be faster than the erosion rate, or else there would be ice at the surface. Additionally, because we did not encounter ground ice at any of the sites, the sublimation rate needs to be fast enough that the ice front is below the depth of our lowest sample. The easiest way to enforce this is to ensure that the t_{till} for all samples is > 0:

$$t_{till} = t_f - \frac{(z_j + \varepsilon_2 \cdot t_f) \cdot (1 - C_0) \cdot \rho_{ice}}{C_0 \cdot \rho_{rock} \cdot \varepsilon_1} > 0 \quad . \tag{4}$$

For the inherited nuclides, we need to make sure that the ratio of ^{26}Al to ^{10}Be lies within the permissible region [*Klein et al.*, 1986], so the final constraint is that:

$$\frac{1}{\lambda_{10}} \ln\left(1 - \frac{N_{10,inh}\lambda_{10}}{P_{10}}\right) \leq \frac{1}{\lambda_{26}} \ln\left(1 - \frac{N_{26,inh}\lambda_{26}}{P_{26}}\right) . \tag{5}$$

The constraint on the inherited nuclides described by equation (5) is applied to both the erosion and sublimation exposure models.

To solve for the unknowns in equations (1) and (2), we find the combination of parameter values that produce the least difference between the observed and modeled concentration profiles and satisfy all the above mentioned conditions. We compare this prediction with the measured nuclide concentrations using an error-weighted chi-square fit and we minimize the chi-square fit using standard optimization techniques in the MATLAB® software. Optimizing the parameters for the lowest chi-square value can be challenging because equations (1) and (2) are nonlinear and asymptotic in behavior for large time values, but avoiding local minima solutions can be addressed by using appropriate starting values for the parameters in the optimization process. To assess the uncertainties of the best-fit parameters for the exposure model, we carry out a 10,000 run Monte Carlo simulation that takes into account uncertainties in the measured nuclide concentrations.

3.4. Results

Figure 3.4 shows the nuclide concentrations for each pit plotted with a line representing the expected nuclide concentrations for an erosion rate that matches the concentration of the uppermost sample in each till. For a stable, or steadily eroding till with no vertical mixing (equation 1), the predicted nuclide concentrations should fall along a curve with an e-folding length equal to the attenuation length, 150 g cm^{-2}. Inherited nuclides, due to prior exposure before the deposition of the till, will increase the concentration of cosmogenic nuclides in the samples, and the resulting expected nuclide concentration curve will be steeper and decrease less rapidly with shielding mass than if no inherited nuclides remain in the samples today. In contrast, the sublimation and erosion exposure model (equation 2) predicts lower nuclide concentrations at depth than equation (1) because the lower samples have spent more time in ice than the upper samples, and have subsequently been shielded from the cosmic ray flux by more mass [*Schäfer et al.*, 2000; *Marchant et al.*, 2002; *Ng et al.*, 2005].

The lower samples in Pit 15 (Beacon Valley moraine) and the Quartermain I till samples at Pit 6 have lower nuclide concentrations than expected for a steadily eroding profile, which is characteristic of a till that formed at least partially as a lag from the sublimation of debris-laden ice. The Arena till samples in the lower section of Pit 6, fall directly on a curve for a steadily eroding profile, which indicates that this unit at

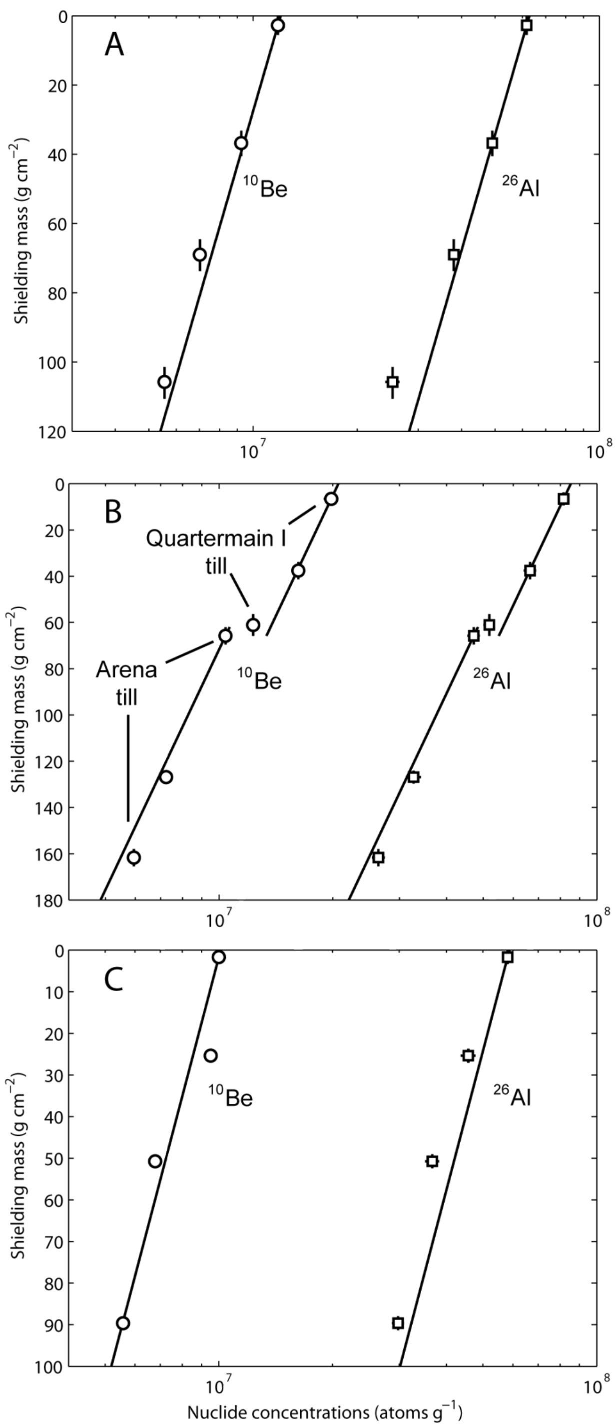

Figure 3.4. Measured ^{10}Be (circles) and ^{26}Al (squares) concentrations (atoms g^{-1}_{quartz}) at each sample site. For each data point, the small vertical lines show the depth range for each sample in the profile, and the small horizontal lines show the 1σ error for each measurement. The solid lines show the predicted nuclide concentration for the erosion rate indicated by the concentration of the surface sample. A) 04-BV-Pit 15: Beacon Valley moraine, B) 04-AV-Pit 6: Quartermain I till overlies Arena till, and C) 04-AV-Pit 16: Arena till.

this site has not had any ice sublimate during the time period that these samples record (~2 Ma). For the Arena till samples at Pit 16, which is where the Arena till crops out at the surface, the ^{26}Al concentrations are lower than predicted by erosion alone, but the ^{10}Be concentrations do not show this pattern. The following sections will explore the results for each site in more detail.

3.4.1. Beacon Valley moraine (04-BV-Pit 15)

Though we do not know exactly when this moraine in Beacon Valley was deposited, we do know that the moraine has experienced some erosion because a desert pavement caps the till abruptly, indicating that the pavement formed as a lag during deflation. Because the deposit is a till, we can assume that it was well-mixed when it was initially deposited, so we will assume that any inherited nuclides the samples contained due to exposure before deposition in the moraine are also well mixed. With no erosion, the measured nuclide concentrations for this till do not yield the same simple exposure age at all depths. The upper two samples have simple exposure ages of ~400 ka, while the lowest sample has an exposure age of ~600 ka. The nuclide concentrations at this site are not well explained by erosion and are characteristic of a till that formed by sublimation of debris-laden ice (Figure 3.4A).

The best-fit results for equation (2) are that the moraine was deposited 0.81 -0.11/+0.12 Ma ago as a deposit of dirty ice with a debris concentration of 7.7 -1.3/+1.0%. This ice sublimed at a rate of 12 ± 1.6 m Ma^{-1} and the till that formed as a result of the sublimation eroded at a rate of 0.78 ± 0.049 m Ma^{-1}. Under these degradation rates, the lowest sample at this site emerged from the ice 0.13 Ma ago. The inherited nuclide concentrations remaining in the till today are 4.4 ± 0.17 × 10^{6} atoms g^{-1} for ^{10}Be and 19 -0.99/+0.42 × 10^{6} atoms g^{-1} for ^{26}Al. These parameters fit the measured concentrations with a chi-square value of 4.1, which is a significant improvement over the erosion exposure model (erosion rate = 1 m Ma^{-1} for the past 3 Ma, chi-square value of 27). The predicted nuclide concentrations for both exposure models are shown in Figure 3.5.

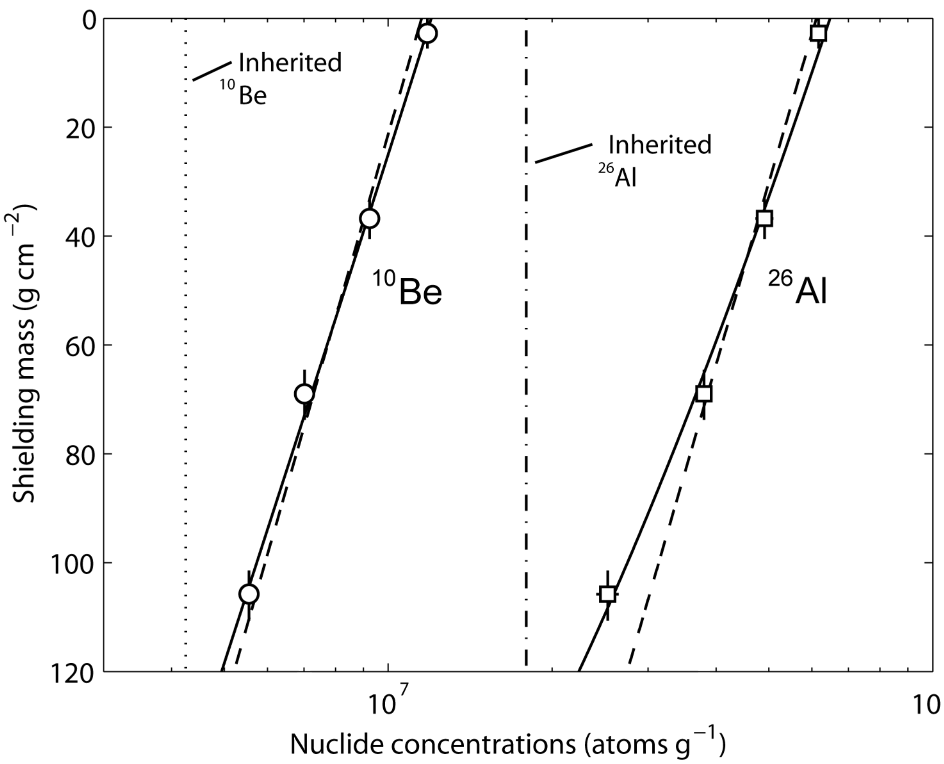

Figure 3.5. Nuclide concentrations for 04-BV-Pit 15, the Beacon Valley moraine, showing the best-fit predictions for the sublimation exposure model (solid line) and the erosion exposure model (dashed line). The dashed and dot-dashed vertical lines show the best-fit inherited ^{10}Be and ^{26}Al, respectively, for the sublimation exposure model. The sublimation exposure model captures the nuclide concentrations at all depths, but the erosion exposure model fails to explain the measured nuclide concentrations. The model parameters for the best-fit sublimation exposure model are: 12 m Ma^{-1} for the sublimation rate, 0.78 m Ma^{-1} for the erosion rate, 7.7% for the initial debris concentration, 0.81 Ma for the age of the moraine, and inherited nuclide concentrations remaining in the till today of 4.4 × 10^{6} atoms g^{-1} for ^{10}Be and 19 × 10^{6} atoms g^{-1} for ^{26}Al.

Though this age for the moraine is consistent with the range of ages for the Taylor moraines in Arena Valley (0.12 – 2.2 Ma), we cannot confidently determine a unique age for this moraine because the nuclide concentrations have reached equilibrium with the sublimation rate and only reflect the degradation rates. Nuclide concentrations will come into equilibrium with a degradation rate once 3 or 4 effective half-lives of the nuclide have passed, which is given by $\tau_{1/2,eff} = \ln(2) / (\lambda_i + \varepsilon/\Lambda)$. Under a sublimation rate of 12 m Ma^{-1} the effective half-life of ^{10}Be and ^{26}Al is on the order of 10^5 years, so the nuclide concentrations will reflect only the sublimation rate and not the age of the moraine once ~0.4 Ma have passed. Implications of this result will be addressed in the Discussion section.

3.4.2. Quartermain till over Arena till (04-AV-Pit 6)

This site presents an unusual geologic scenario where the Quartermain I till overlies the Arena till, and an undisturbed, *in-situ* desert pavement separates the two units. From the geologic setting, we know that the Arena till was deposited first, and was subject to surface erosion because the desert pavement caps the unit abruptly. Then the Arena till was buried for some time by the Taylor Glacier that deposited the Quartermain I till on top of the desert pavement that caps the Arena till. The Quartermain I till is itself capped by a desert pavement, which indicates that this unit has undergone erosion too. Brook et al. [1995] dated boulders from the Quartermain I till using cosmogenic nuclides and updated previously dated boulders [*Brown et al.*, 1991; *Brook et al.*, 1993] and found that the oldest boulder had a minimum age of 3.5 Ma, which we will use as a minimum age for the unit in our analysis. Because both deposits are tills, we will consider that any inherited nuclides are well-mixed throughout the unit, and we will allow for different inherited nuclide concentrations in each till.

Using the exposure model with only erosion, equation (1), the erosion rate that best fits the nuclide concentrations is 0.59 m Ma^{-1}, but the chi-square fit is 29. We fit a single erosion rate to both tills because the limiting age of the Quartermain I till (> 3.5

Ma) is old enough that nuclide concentrations in both units should have come into equilibrium with the more recent surface erosion rate. The erosion exposure model fits the samples from the Arena till very well, but the samples in the Quartermain I till have lower nuclide concentrations than predicted by erosion alone (Figure 3.4B), which suggests that the sublimation and erosion exposure model may be more appropriate for the Quartermain I till samples. The parameters for equation (2) (Figure 3.6) that best match the measured nuclide concentrations for the Quartermain I till samples are that the till was deposited at 5.3 -0.45/+0.33 Ma ago as a deposit of dirty ice with a debris concentration of 33 ± 8.4%. This ice sublimed at a rate of 0.78 -0.28/+0.16 m Ma^{-1} and the resulting till eroded at a rate of 0.50 -0.06/+0.073 m Ma^{-1}. Under these degradation rates, the lowest sample at this site emerged from the ice at 0.27 Ma ago. The inherited nuclides remaining in the samples today in this scenario are zero for ^{26}Al and $2.6 \times 10^6 \pm 7.6 \times 10^3$ atoms g^{-1} for ^{10}Be. This age for the till is older than 3.5 Ma, but the nuclide concentrations have reached equilibrium with the degradation rates, so the age cannot be uniquely determined. These parameters fit the measured concentrations with a chi-square value of 4.0.

The Arena till data fit an erosion line very well and do not show the characteristic pattern of sublimation, so it does not appear that there has been any ground ice in the underlying Arena till during the period of time that these samples record (~2.4 Ma). We can model the predicted nuclide concentrations for the Arena till samples if they were subject to the change in shielding resulting from the degradation rates that best-fit the Quartermain till nuclide concentrations, and not include any ice in the Arena samples. With these conditions, the parameters that fit the Quartermain I till fit the Arena till data well (chi-square = 2.7) with inherited nuclides of $7.9 \pm 1.4 \times 10^5$ atoms g^{-1} for ^{10}Be and a value that is indistinguishable from zero ($5.5 \pm 13 \times 10^4$ atoms g^{-1}) for ^{26}Al remaining in the underlying Arena till today (Figure 3.6). The remaining inherited nuclide concentrations in the Arena till today were acquired when the Arena till was at the surface before the site was buried by ice and the Quartermain I till was deposited, and these results are consistent with the age of the Quartermain I till. This

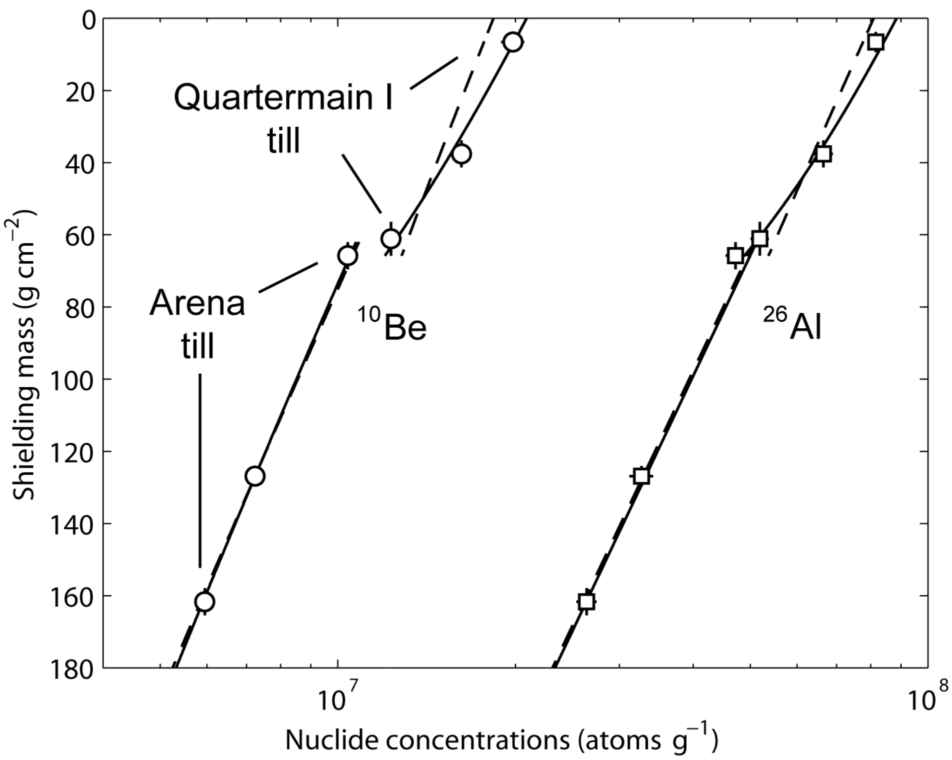

Figure 3.6. Nuclide concentrations for 04-AV-Pit 6, the site where the Quartermain I till overlies the Arena till, showing the best-fit predictions for the sublimation exposure model (solid line) and the erosion exposure model (dashed line). For the Quartermain I till, the sublimation exposure model captures the nuclide concentrations much better than the erosion model. The model parameters for the best-fit sublimation exposure model are 0.78 m Ma^{-1} for the sublimation rate, 0.50 m Ma^{-1} for the erosion rate, 33% for the initial debris concentration, 5.3 Ma for the age of the till, and inherited nuclide concentrations remaining in the till today of 2.6×10^6 atoms g^{-1} for ^{10}Be (too small to plot at this scale) and essentially zero for ^{26}Al. For the lower three samples from the Arena till, these degradation rates predict nuclide concentrations shown by the solid lines with 7.9×10^5 atoms g^{-1} for inherited ^{10}Be that was acquired before the Quartermain I till was deposited remaining in the till today.

exercise demonstrates that the Arena till data are consistent with the sublimation and erosion rates from the overlying Quartermain till, without having any ice in the Arena till.

3.4.3. Arena till (04-AV-Pit 16)

At this site, the Arena till crops out at the surface and is not covered by the Quartermain I till. It is not clear if the Quartermain I till never covered this site, or if erosion has removed it, but it is clear that the site has experienced erosion because the till is abruptly capped by a desert pavement. Based on similar soil properties and stratigraphic relationships with other deposits, Marchant et al. [1993a] determined that the Arena till is >11.3 Ma. Given the great age of this deposit, there should be no inherited nuclides from exposure before the till was laid down remaining in the samples today, but we will test for the signature of inherited nuclides in the data with the exposure models. The measured nuclide concentrations for this site yield simple exposure ages of 400 – 600 ka, which indicates that the nuclide concentrations reflect the degradation rate at the site and not the exposure age.

The parameters for equation (1) that best-fit the measured nuclide concentrations are that there is no inherited ^{10}Be or ^{26}Al remaining in the sample today and that the site has been eroding at a rate of 1.2 ± 0.025 m Ma^{-1} (Figure 3.7). Though the ^{26}Al data show lower nuclide concentrations than expected by the erosion line, which is characteristic of a till formed by the sublimation of ice, the ^{10}Be data do not show this same pattern (Figure 3.4C), and the sublimation exposure model does not improve the fit to the data (Figure 3.7). Thus, the best-fit results for this site are that the Arena till has been eroding at a rate of 1.2 m Ma^{-1} for the past 1.5 Ma, which is the length of time it takes for the ^{10}Be concentrations to equilibrate with this erosion rate.

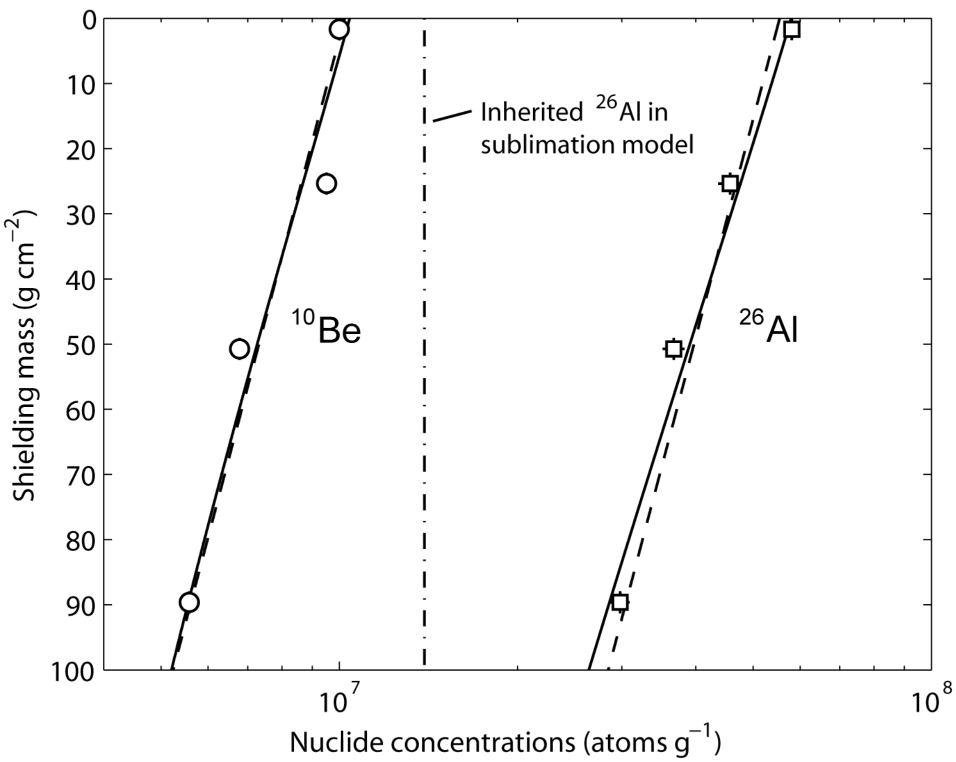

Figure 3.7. Nuclide concentrations for 04-AV-Pit 16, the Arena till, showing the best-fit predictions for the sublimation exposure model (solid line) and the erosion exposure model (dashed line). The dot-dashed vertical line shows the best-fit inherited ^{26}Al for the sublimation exposure model, and the inherited ^{10}Be (3.3×10^6 atoms g^{-1}) is too small to plot on this scale. At this site, the sublimation exposure model does not explain the data any better than the erosion exposure model, and the best-fit results is that this site has been eroding at 1.2 m Ma^{-1} for the past 1.5 Ma, with no inherited nuclides remaining in the till today.

3.5. Discussion

Though ground ice was not found at any of the sample sites in this study, two of the three units that were sampled have cosmogenic nuclide concentration patterns consistent with a till that formed in part as a lag by the sublimation of debris-laden ice. This finding is consistent with our observation and that made by others that many modern and recently formed moraines in the MDV are ice-cored [*Debenham*, 1921; *Hall et al.*, 1993; *Denton and Marchant*, 2000], and that till patches on valley floors can form by sublimation of debris-laden ice [*Sugden et al.*, 1995]. The degradation rates that best fit the measured nuclide concentrations range from 0.7 – 12 m Ma^{-1} for the sublimation of ice and the resulting till eroded at a rate of 0.5 – 1.2 m Ma^{-1}. Previous studies using cosmogenic nuclides to determine ice sublimation rates through the Granite Drift in Beacon Valley calculated rates from 4 – 23 m Ma^{-1} [*Schäfer et al.*, 2000; *Ng et al.*, 2005], and regolith erosion rates in Arena Valley from 0.2 – 2 m Ma^{-1} [*Putkonen et al.*, 2008a; *Morgan et al.*, 2008].

The time period that these nuclide concentrations record is limited by the effective half-life of the samples, which is dependent on the degradation rate at each site. Once several effective half lives have passed, the cosmogenic nuclide concentration in a given sample will reflect only the degradation rate, and not the exposure age of the sample. At erosion rates of 0.5 – 1.2 m Ma^{-1}, the effective half life of ^{26}Al is 0.2 – 0.4 Ma and for ^{10}Be it is 0.3 – 0.6 Ma. Once ice is no longer shielding the lower samples, it will take ~1 – 1.5 Ma for the ^{26}Al to reach equilibrium and ~1.2 – 2.4 Ma for the ^{10}Be to reach equilibrium with these erosion rates. This fact is what limits our ability to uniquely define an exposure age for these glacial deposits and limits our findings to the past 1-2 Ma. Because the nuclide concentrations of the lowest samples from the Beacon Valley moraine and Quartermain I tills are lower than the erosion rate set by the surface samples, these lowest samples have not been shielded by ice at some point in the last ~1-2 Ma. If these samples had accreted into the till > ~2 Ma, then their nuclide concentrations would reflect only the surface erosion rate. Though we cannot uniquely determine the depositional age of either of these deposits, the results of the Monte Carlo simulation show that the lowest sample at the Beacon Valley moraine accreted into the till at 65 ka, and at the Quartermain I till the lowest sample emerged from ice at 270 ka.

At the Beacon Valley moraine site (Pit 15) we can constrain the model results with the geologic setting because this site is at the crest of a moraine that has a width between 10 – 25 m from the crest to the toe of the southeastern flank. In a cross-sectional profile, moraines are often modeled as having initial slopes at the angle of repose (~34°) with a triangular cross-section [*Hallet and Putkonen*, 1994; *Putkonen and Swanson*, 2003; *Putkonen and O'Neal*, 2006; *Putkonen et al.*, 2008b]. At the sample site today, the moraine is 20 m wide and 1.8 m high, giving it a slope of 5°. If we assume that the maximum height of the moraine is limited by the angle of repose, and that the moraine was initially 20 m wide, then the initial moraine height would have to be < 13.5 m, otherwise the slope would have exceeded the angle of repose. The exposure model results suggest that over 0.81 Ma, 9.5 m of ice and 0.63 m of till have been removed from this site, which would give the moraine an initial height of ~12 m and an initial slope of 31° at the sample site. This result is consistent with the suggestion that the effect of the ice-core in the moraine is that the moraine will start out with a lower angle than the angle of repose [*Putkonen et al.*, 2008b]. This analysis limits the initial height of the moraine, which also allows us to constrain the age of the moraine to < 1.1 Ma old, otherwise the moraine would have eroded to a lower height than it has today, and suggests that this moraine correlates with the Taylor IVa (1.0 Ma) moraine in Arena Valley.

The results from the sublimation exposure model suggests that the Beacon Valley moraine and Quartermain I tills began as dirty ice deposits with debris concentrations between 7.9 – 33%. In comparison, the debris concentration of the buried relict glacial ice in Beacon Valley is 3.1% by volume, and this debris consists of 27% gravel, 54% sand, and 17% mud [*Marchant et al.*, 2002]. Other samples of relict glacier ice in Beacon Valley have debris concentrations of ~10% [J. Stone, personal communication, 2009]. The result that the Beacon Valley moraine began as an ice-cored moraine with ~8% debris in it is consistent with the debris concentration in the relict glacier ice in Beacon Valley. At the Quartermain I site, our results suggest that the initial debris concentration is 33 ± 8.4%, which is approaching a concentration where the deposit would no longer be ice-supported. The sublimation exposure model cannot distinguish the source of ice in the deposit, and because the Quartermain I till is quite old (> 3.5 Ma) and we can only say that the deposit had ice in it as recently as 270 ka, it is possible that the ice in this till could be ground ice and not relict glacier ice that was emplaced at the same time the till was deposited.

Though the model results suggest that ice was present at two of the sites, there does not appear to be any overturning of the till due to cryoturbation during the last 1-2 Ma. This is consistent with the lack of geomorphic features that would indicate cryoturbation at the sites (e.g. polygons, sand wedges) and the cosmogenic profiles over the Granite Drift [*Schäfer et al.*, 2000; *Marchant et al.*, 2002]. These data are consistent with the idea that erosion in the MDV is limited to the upper few centimeters of the surface, and does not involve creep or vertical mixing [*Putkonen et al.*, 2008a]. Because erosion in the MDV appears to occur without creep or vertical mixing, the surface appears to simply lower over time. This may explain how surface deposits in the MDV maintain their meter-scale morphology for millions of years while uniformly eroding a few meters over that time period.

Erosion of the till layer that caps relict glacier ice or that contains ground ice may partially explain why ice is found close to the surface in millions of years old deposits despite the fact that measured and modeled sublimation rates suggest that these deposits should be free of ice. If erosion is lowering the surface at a rate close to that at which sublimation is lowering the ice front, then ground ice or even relict glacial ice may persist near the surface. The presence of relict glacier ice below a 50 cm thick layer of sublimation till has been interpreted as an indication that the till layer protects the underlying ice from sublimation, and that the sublimation rate through this till layer is very low [*Sugden et al.*, 1995; *Marchant et al.*, 2002]. Our results suggest that the relict glacier ice in Beacon Valley may only be covered by a thin layer of till because this till layer is continuously eroding.

3.6. Conclusions

We addressed the apparent contradiction between millions of years old glacial deposits that contain ground ice and maintain their meter-scale morphology despite the fact that measured and modeled degradation rates suggest that these deposits should be ice-free and have noticeably eroded over these time periods. This was addressed by analyzing the concentration of cosmogenic ^{10}Be and ^{26}Al in soil profiles from three glacial deposits in the Quartermain Mountains, Antarctica. We found that the measured nuclide concentrations are inconsistent with the published ages of the deposits, that erosion alone does not always explain the discrepancies in concentrations, and that even though we did not encounter ice during the excavation of these deposits, nuclide concentrations in Beacon Valley moraine and the Quartermain I till are best explained by the sublimation of debris laden ice. The degradation rates that best match the data range from 0.7 – 12 m Ma^{-1} for the sublimation of ice with initial debris concentrations ranging from 7.9 – 33% and erosion of the overlying till at rates of 0.5 – 1.2 m Ma^{-1}. Because nuclide concentrations will reach equilibrium with these degradation rates on the order of 0.4 – 2 Ma, we cannot uniquely determine the age of these deposits, but we can establish that the lowest samples in the Beacon Valley moraine and the Quartermain I till emerged from ice as recently as 65 ka and 270 ka, respectively.

The model results suggest that the Beacon Valley moraine started out as an ice-cored moraine with a debris concentration of ~8%, which is similar to that of the buried, relict glacier ice in Beacon Valley. Though the measured nuclide concentrations reflect only the degradation rate of the moraine, the best-fit age and geomorphic setting suggest that this moraine corresponds with the Taylor IVa moraine (1.0 Ma) in Arena Valley. At the Quartermain I till site, the modeled initial debris concentration is much higher (33 ± 8.4%), and may indicate that the ice at this site is ground ice, and not relict glacial ice. The degradation rates of the Quartermain I till only allow us to record what happened at this site for the last 1-2 Ma, so it is possible that this site reflects the growth, and subsequent sublimation, of ground ice in the unit.

Our results show that some glacial deposits in the MDV have evolved due to both the sublimation of ground ice and by erosion of the surface. The tills where the nuclide concentrations are characteristic of the sublimation of ice have degraded more due to sublimation than by erosion, which suggests that some deposits may be more modified by this process than by active erosion at the surface. Degradation by the sublimation of either massive or ground ice in glacial deposits, coupled with the observation that this degradation occurs without vertical mixing or overturning of the soil, may be part of the reason that glacial landforms such as moraines in the Quartermain Mountains stand in sharp relief even though they are millions of years old. As ice sublimes through a till layer the permafrost table will lower through a soil, and if the overlying till is also lowering due to erosion then the permafrost table may persist near the surface. Erosion of the units above ground ice or relict glacier ice may explain how this ice continues to lie beneath a thin layer of overlying debris even though it is actively subliming, without a need to invoke recharge mechanisms.

Chapter 4

CONCLUSIONS

The focus of this dissertation is the evolution of unconsolidated deposits in the McMurdo Dry Valleys, Antarctica (MDV) over millions of years time periods. Using the concentration of cosmogenic nuclides ^{10}Be and ^{26}Al from vertical profiles in meter-deep hand-dug soil pits, I established degradation rates and limiting ages for unconsolidated deposits in the MDV. By quantifying rates of geomorphic processes that operate over long time periods in the MDV I provide a clearer picture of how these landscapes are actively changing over millions of years. This chapter summarizes these results and presents directions for future work.

4.1 Summary of results

In Chapter 2, I explore the interpretation that ash layers found throughout the MDV indicate slope stability by measuring the concentration of ^{10}Be and ^{26}Al from colluvium deposits below two well-studied ash layers, the Arena Valley ash (^{40}Ar/^{39}Ar age of 4.33 Ma [*Marchant et al.*, 1993a]) and the Hart ash (K-Ar age of 3.9 Ma [*Hall et al.*, 1993]), as well as from sands intermixed with the upper layer of the ashes. Commonly, ash deposits at or near the surface are interpreted as indicating extremely low slope evolution at that site [*e.g. Marchant et al.*, 1996]. At both of the ash sites, the measured concentrations of ^{10}Be and ^{26}Al are significantly lower than expected given the age of the *in situ* airfall ashes, and burial by ice or soil for a long period of time cannot explain these low concentrations. The nuclide concentrations are consistent with the depositional ages of the ashes only if they are interpreted as the result of degradation of regolith or ash above the present-day ash surface at these sites. The best-fit erosion rates for the nuclide concentrations are $3.5 \pm 0.41 \times 10^{-5}$ g cm^{-2} yr^{-1} (0.19 m Ma^{-1} of regolith or 0.35 m Ma^{-1} of ash) for the past ~4.33 Ma at the Arena Valley ash site, and $4.8 \pm 0.21 \times 10^{-4}$ g cm^{-2} yr^{-1} (2.6 m Ma^{-1} of regolith or 4.7 m Ma^{-1}

of ash) for the past ~1 Ma at the Hart ash site. While these rates are slow, they are measurable, and given an additional 0.1 – 0.5 Ma the ash deposits at these sites would be entirely removed from these sites. At both of these sites there is no indication of soil creep or vertical mixing of the sediment in the last ~1 – 4.33 Ma. The cosmogenic nuclide concentrations in the samples above the ash do not have the same exposure history as those below the ash, supporting the idea that regolith degradation in the MDV is limited to the upper few centimeters of the regolith [*Putkonen et al.*, 2008a]. These results are consistent with the idea that geomorphic activity slows with increased elevation and distance from the coast in the MDV [e.g. *Marchant and Head*, 2007]. Because erosion above the present-day ash surface has occurred, there are disconformities at these sites, indicating that part of the geologic record has been lost.

In Chapter 3, I address the apparent contradiction between millions of years old glacial deposits that contain ground ice and maintain their meter-scale morphology despite the fact that measured and modeled degradation rates suggest that these deposits should be ice-free and have noticeably eroded over these time periods. By analyzing the concentration of cosmogenic ^{10}Be and ^{26}Al in soil profiles from three glacial deposits in the Quartermain Mountains, Antarctica, I found that the measured nuclide concentrations are inconsistent with the ages of the deposits, that erosion alone does not always explain these concentrations, and that even though I did not encounter ice during the excavation of these deposits, nuclide concentrations in Beacon Valley moraine and the Quartermain I till are best explained by the sublimation of ice. The degradation rates that best match the data range from 0.7 – 12 m Ma^{-1} for the sublimation of ice with initial debris concentrations ranging from 7.9 – 33% and erosion of the overlying till at rates of 0.5 – 1.2 m Ma^{-1}. Overturning of the tills by cryoturbation, vertical mixing, and soil creep is not indicated by the cosmogenic nuclide profiles, and degradation appears to be limited to within a few centimeters of the surface. Erosion of these tills without vertical mixing may explain how glacial deposits in the MDV maintain their morphology and contain ground ice for millions of years.

Collectively these results paint a picture of a more dynamic environment in the MDV than previously thought and suggest that a sizeable fraction of the sedimentary record may have been remobilized. Erosion rates of 0.2 – 4.7 m Ma^{-1} determined in this study are in fact higher than the lowest bedrock erosion rates (0.14 m Ma^{-1}) in the MDV [*Summerfield et al.*, 1999]. Erosion has been active at all of the sites examined in the project, which indicates that some of the sedimentary record of the MDV has been lost at these sites. The deposits where the nuclide concentrations are characteristic of the sublimation of ice have degraded more due to sublimation than by erosion, which suggests that some deposits may be more modified by this process than by active erosion at the surface. Degradation by the sublimation of either massive ice or pore ice in glacial deposits, coupled with the observation that this degradation occurs without vertical mixing or overturning of the soil, may be part of the reason that glacial landforms such as moraines in the MDV stand in sharp relief even though they are millions of years old. These results also support the notion that regolith degradation in the MDV occurs without creep, which is a process that is ubiquitous on the planet [*Oehm and Hallet*, 2005], and that erosion is limited to the upper few centimeters of the regolith.

Because all of the sites studied in this project have cosmogenic nuclide concentration profiles that decrease monotonically with increasing soil depth vertical mixing of the regolith is not indicated as an active geomorphic process in the timeframe that the nuclide concentrations record (~1 – 4.33 Ma). If vertical mixing was active during this time period, the nuclide concentration profiles would be expected to be well-mixed and would not fit an exponential profile.

If soil creep were to occur in a laminar manner and there was no vertical mixing of the regolith as it moved downhill, then the nuclide concentrations would still be expected to fall along an exponential depth profile because the shielding mass of the samples would not change through time. However, the unconsolidated deposits studied in this project are all poorly-sorted glacial tills and colluvium where internal shear across large boulders would necessarily disturb the soil profile. Therefore laminar soil creep is highly unlikely to occur at this site.

Plug flow, where all the shear in the soil profile is concentrated in a narrow depth zone at the base of the moving layer, would also maintain an unmixed, exponential isotope concentration profile in the soil. However, such flows are typically related to coherent landslides or gelifluction, which both leave distinct scars, steps, and terraces in the landscape. No such features are detected in the field area suggesting that plug flows, even if present, are rare.

The active aeolian and surface flux of material across the MDV is well documented[*e.g. Lancaster*, 2002; *Putkonen et al.*, 2007], but the results of this project show that any downhill transport of mineral material must be limited to the upper few centimeters because creep is not indicated by the cosmogenic isotope concentrations that were measured just below the surface. Because soil creep rates are closely related to the rates of landscape evolution [*Oehm and Hallet*, 2005], the lack of creep at the sample sites partly explains the low erosion rates determined in this study.

The erosion rates determined in this study (0.2 – 4 m Ma^{-1}) are similar to the lowest bedrock erosion rates ever reported, which also come from other arid regions [*Cockburn et al.*, 1999; *Summerfield et al.*, 1999; *Belton et al.*, 2004; *Dunai et al.*, 2005]. Even though the degradation rates reported in this project are very low, they will transform unconsolidated deposits over millions of years, and the MDV should not be considered a relict landscape where no landscape evolution is occurring. However, the degradation rates of unconsolidated deposits must be considered among the lowest rates on Earth, and do imply that we must consider the evolution of landscapes over millions of years timescales. The results of this project show that degradation in the MDV occurs at a steady rate through time, with no vertical mixing or downslope creep of material below the upper few centimeters of the surface. This suggests that slopes should retreat in parallel, which may help to preserve the apparent contradiction between erosion rates and landforms that maintain their meter-scale morphology for millions of years.

4.2 Future work

One of the major findings in this project is that erosion in the MDV appears to occur without any creep or vertical mixing of the regolith, suggesting that any erosion that occurs is limited to a thin layer right at the surface. The lack of vertical creep and mixing is likely related to the lack of moisture in the regolith that could induce frost heave, and because the MDV lack higher lifeforms that would cause bioturbation. Creep is a universal process that has the ability to move regolith downslope at rates on the order of cm yr^{-1} [*Oehm and Hallet*, 2005], and the role of biota in arid geomorphological systems is the subject of much recent work [*Tooth*, 2008 and references therein]. Because unconsolidated deposits in the MDV can be viewed as an end-member in the spectrum of both soil moisture content and biotic activity, further work should be undertaken to quantify the role of biotic activity and soil moisture in inducing the downslope creep of regolith.

Research in landscape evolution over millions of years has recently begun to explore the relationships between tectonics, climate, and erosion, and especially the isostatic responses to changes in the denudation of landscapes [*e.g. Molnar and England*, 1990]. In the MDV, we need to understand the role of climate and tectonics in governing the slow rates of landscape evolution that we quantified in this study. For example, most studies of the interplay between climate, tectonics, and erosion have been done at subduction zones [*e.g. Montgomery et al.*, 2001]. The MDV lie at the edge of a rift margin, and the expected feedbacks between tectonics, climate, and erosion are not well-studied for this tectonic setting. Additionally, the cold, hyperarid climate of the MDV is one of the most extreme climates found on Earth, so the MDV could be viewed as an end-member for the climate spectrum to develop the framework for landscape evolution under these three governing processes. Furthermore, the onset of hyperarid conditions in the MDV remains an unresolved issue, but this timing should have tectonic and erosional implications that should be explored.

It has been suggested that in areas of low humidity and tectonic activity, such as the MDV, that landscape evolution should be considered a ‘directional’ (evolutionary) rather than cyclical or in equilibrium [*Ollier*, 1979]. The results of this study show that the landscape of the MDV experience a slow, but steady lowering of the surface, which supports the notion that landscape evolution over millions of years is directional rather than cyclical. It would be informative for geomorphologists to consider landscape evolution in the MDV on timescales even longer than examined in this study, perhaps with apatite fission track thermochronology, so that we can further explore the feedbacks between climate, tectonics, and erosion that may, or may not be occurring in the unique environment of the MDV.

The lowest bedrock erosion rates on Earth are found in the arid deserts, such as the MDV, the Atacama Desert in Chile, and the Namib Desert in Namibia, or in the stable cratons of continents, such as central Australia. Most studies of landscape evolution in these other deserts have been done on either bedrock outcrops [*e.g. Cockburn et al.*, 1999; *Belton et al.*, 2004] or on boulders from ancient landforms [*e.g. Dunai et al.*, 2005; *Schafer et al.*, 1999]. More work should be undertaken to understand how unconsolidated deposits degrade in these other field areas because it is often the landforms consisting of unconsolidated deposits that contain the paleoclimate information that we infer from landscapes, and it is fundamental to our interpretation of these landforms that we understand how much they have degraded since they were deposited.

Lastly, we must consider the implications that these results have for the glacial history of the EAIS. Because the MDV contain the largest terrestrial record of Neogene ice-sheet history, they have often been considered the best archive of EAIS glacial history. However, there is a recent suggestion that the hyperarid condition in the MDV may be a unique environment, and are not representative of the rest of the continent [*Fox*, 2008]. Undertaking projects similar to this study in other ice-free regions in Antarctica would help us address whether or not the MDV are the appropriate site to examine the glacial history of the EAIS.

REFERENCES

Balco, G., J. Stone, and C. Jennings (2005a), Dating Plio-Pleistocene glacial sediments using the cosmic-ray-produced radionuclides 10Be and 26Al, *American Journal of Science*, *305*(1), 1-41, doi:10.2475/ajs.305.1.1.

Balco, G., J. Stone, and J. Mason (2005b), Numerical ages for Plio-Pleistocene glacial sediment sequences by 26Al/10Be dating of quartz in buried paleosols, *Earth and Planetary Science Letters*, *232*(1-2), 179-191.

Barrett, P. J. (1981), History of the Ross Sea region during the deposition of the Beacon Supergroup 400-180 million years ago, *Journal of the Royal Society of New Zealand*, *11*, 447-458.

Barrett, P. J., C. J. Adams, W. C. Mclntosh, C. C. Swisher, and G. S. Wilson (1992), Geochronological evidence supporting Antarctic deglaciation three million years ago, *Nature*, *359*(6398), 816-818, doi:10.1038/359816a0.

Barrett, P. J., and M. J. Hambrey (1992), Plio-Pleistocene sedimentation in Ferrar Fiord, Antarctica, *Sedimentology*, *39*(1), 109-123.

Bates, R. L., and J. A. Jackson (1984), *Dictionary of Geological Terms*, Third Edition., Anchor Books, New York.

Belton, D. X., R. W. Brown, B. P. Kohn, D. Fink, and K. A. Farley (2004), Quantitative resolution of the debate over antiquity of the central Australian landscape: implications for the tectonic and geomorphic stability of cratonic interiors, *Earth and Planetary Science Letters*, *219*(1-2), 21-34.

Bierman, P. R., and K. K. Nichols (2004), Rock to Sediment - Slope to Sea with 10Be - Rates of Landscape Change, *Annual Review of Earth and Planetary Sciences*, *32*, 215-255, doi:10.1146/annurev.earth.32.101802.120539.

Bockheim, J. G. (2002), Landform and Soil Development in the McMurdo Dry Valleys, Antarctica: A Regional Synthesis, *Arctic, Antarctic, and Alpine Research*, *34*(3), 308-317.

Bockheim, J. (2007), Soil processes and development rates in the Quartermain Mountains, Upper Taylor Glacier Region, Antarctica, *Geografiska Annaler, Series A: Physical Geography*, *89*(3), 153-165, doi:doi:10.1111/j.1468-0459.2007.00315.x.

Bockheim, J., and R. Ackert Jr. (2007), Implications of soils on mid-Miocene-aged drifts in the McMurdo Dry Valleys for ice sheet history and paleoclimate reconstruction, *Geomorphology*, *92*(1-2), 12-24.

Bockheim, J., I. Campbell, and M. McLeod (2007), Permafrost Distribution and Active-Layer Depths in the McMurdo Dry Valleys, Antarctica, *Permafrost and Periglacial Processes*, (18), 217-227, doi:10.1002/ppp.588.

Brook, E., E. Brown, M. Kurz, R. Ackert, G. Raisbeck, and F. Yiou (1995), Constraints on age, erosion, and uplift of Neogene glacial deposits in the Transantarctic Mountains determined from in situ cosmogenic 10 Be and 26 Al, *Geology*, *23*(12), 1063-1066.

Brook, E., M. Kurz, R. Ackert, G. Denton, E. Brown, G. Raisbeck, and F. Yiou (1993), Chronology of Taylor Glacier Advances in Arena Valley, Antarctica, Using in Situ Cosmogenic 3He and 10Be, *Quaternary Research*, *39*(1), 11-23.

Brown, E., J. Edmond, G. Raisbeck, F. Yiou, M. Kurz, and E. Brook (1991), Examination of surface exposure ages of Antarctic moraines using in situ produced Be-10 and Al-26, *Geochimica et Cosmochimica Acta*, *55*(8), 2269-2284.

Calkin, P. (1964), *Geomorphology and glacial geology of the Victoria Valley system, southern Victoria Land, Antarctica.*, Ohio State University Research Foundation, Columbus, Ohio.

Campbell, I. B., and G. G. C. Claridge (1969), A classification of frigic soils-the zonal soils of the Antarctic continent, *Soil Science*, *107*, 75-85.

Campbell, I. B., and G. G. C. Claridge (1987), *Antarctica: Soils, Weathering Processes, and Environment*, Elsevier, New York.

Campbell, I. B., and G. G. C. Claridge (2006), Permafrost properties, patterns and processes in the Transantarctic Mountains region, *Permafrost and Periglacial Processes*, *17*, 215-232.

Chinn, T. J. (1980), Glacier Balances in the Dry Valleys Area, Victoria Land, Antarctica, *Proceedings of the Workshop at Riederalp, Switzerland IAHS-AISH Publication*, (126), 237-247.

Chinn, T. J. (1990), The dry valleys in Antarctica: The Ross Sea region, *Department of Scientific and Industrial Research, Wellington, NZ*, 137-153.

Clow, G. D., C. P. McKay, G. M. Simmons Jr, and R. A. Wharton Jr (1988), Climatological observations and predicted sublimation rates at Lake Hoare, Antarctica, *Journal of Climate*, *1*(7), 715-728.

Cockburn, H. A. P., M. A. Seidl, and M. A. Summerfield (1999), Quantifying denudation rates on inselbergs in the central Namib Desert using in situ-produced cosmogenic 10 Be and 26 Al, *Geology*, *27*(5), 399-402.

Cuffey, K. M., H. Conway, A. M. Gades, B. Hallet, R. Lorrain, J. P. Severinghaus, E. J. Steig, B. Vaughn, and J. W. C. White (2000), Entrainment at cold glacier beds, *Geology*, *28*(4), 351-354, doi:10.1130/0091-7613(2000)28<351:EACGB>2.0.CO;2.

Davis, W. M. (1899), The geographical cycle, *Geographical Journal*, *14*, 481-504.

Debenham, F. (1921), Recent and Local Deposits of McMurdo Sound Region: British Antarctic (Terra Nova) Expedition (1910), *Natural History Report, Geology*, *1*(3), 63-90.

Denton, G. H., and D. R. Marchant (2000), The Geologic Basis for a Reconstruction of a Grounded Ice Sheet in Mcmurdo sound, Antarctica, at the Last Glacial Maximum, *Geografiska Annaler. Series A, Physical Geography*, *82*(2/3, Glacial and Paleoclimatic History of the Ross Ice Drainage System of Antarctica), 167-211.

Denton, G., M. Prentice, D. Kellogg, and T. Kellogg (1984), Late Tertiary history of the Antarctic ice sheet; evidence form the dry valleys, *Geology*, *12*(5), 263-267.

Denton, G., D. Sugden, D. Marchant, B. Hall, and T. Wilch (1993), East Antarctic Ice Sheet Sensitivity to Pliocene Climatic Change from a Dry Valleys Perspective, *Geografiska Annaler. Series A, Physical Geography*, *75*(4, A Special Volume Arising from the Vega Symposium: The Case for a Stable East Antarctic Ice Sheet), 155-204.

Doran, P., C. McKay, G. Clow, G. Dana, A. Fountain, T. Nylen, and W. Lyons (2002), Valley floor climate observations from the McMurdo dry valleys, Antarctica, 1986–2000, *Journal of Geophysical Research*, *107*(D24), 4772, doi:10.1029/2001JD002045.

Drewry, D. J. (1983), The surface of the Antarctic ice sheet, *Antarctica: Glaciological and Geophysical Folio*.

Dunai, T. J., G. A. G. Lopez, and J. Juez-Larre (2005), Oligocene-Miocene age of aridity in the Atacama Desert revealed by exposure dating of erosion-sensitive landforms, *Geology*, *33*(4), 321-324.

Fernandes, N., and W. Dietrich (1997), Hillslope Evolution by Diffusive Processes: The Timescale for Equilibrium Adjustments, *Water Resources Research*, *33*(6), 1307–1318.

Fountain, A. et al. (1999), Physical Controls on the Taylor Valley Ecosystem, Antarctica, *BioScience*, *49*(12, McMurdo Dry Valleys), 961-971.

Fox, D. (2008), Freeze-Dried Findings Support a Tale of Two Ancient Climates, *Science*, *320*(5880), 1152-1154.

Gilbert, G. K. (1909), *The Convexity of Hilltops.*, University of Chicago Press.

Gosse, J., and F. Phillips (2001), Terrestrial in situ cosmogenic nuclides: theory and application, *Quaternary Science Reviews*, *20*(14), 1475-1560.

Hack, J. T. (1975), Dynamic equilibrium and landscape evolution, *Theories of landform development*, 87–102.

Hagedorn, B., R. Sletten, and B. Hallet (2007), Sublimation and ice condensation in hyperarid soils: Modeling results using field data from Victoria Valley, Antarctica, *Journal of Geophysical Research*, *112*, F03017, doi:10.1029/2006JF000580.

Hall, B. L., G. H. Denton, D. R. Lux, and C. Schluchter (1997), Pliocene Paleoenvironment and Antarctic Ice Sheet Behavior: Evidence From Wright Valley, *The Journal of Geology*, *105*(3), 285-294.

Hall, B., G. Denton, D. Lux, and J. Bockheim (1993), Late Tertiary Antarctic Paleoclimate and Ice-Sheet Dynamics Inferred from Surficial Deposits in Wright Valley, *Geografiska Annaler. Series A, Physical Geography*, *75*(4, A Special Volume Arising from the Vega Symposium: The Case for a Stable East Antarctic Ice Sheet), 239-267.

Hallet, B., L. Hunter, and J. Bogen (1996), Rates of erosion and sediment evacuation by glaciers: A review of field data and their implications, *Global and Planetary Change*, *12*(1-4), 213-235.

Hallet, B., and J. Putkonen (1994), Surface Dating of Dynamic Landforms: Young Boulders on Aging Moraines, , 937-940, doi:10.1126/science.265.5174.937.

Hanks, T. C. (2000), The age of scarplike landforms from diffusion-equation analysis, *Quaternary Geochronology: methods and applications (Stratton Noller, J.; Sowers, JM; Lettis, WR; editors). American Geophysical Union*, *582*.

Hanks, T., R. Bucknam, K. Lajoie, and R. Wallace (1984), Modification of wave-cut and faulting-controlled landforms, *Journal of Geophysical Research*, *89*, 5771-5790.

Harwood, D. M., and P. N. Webb (1998), Glacial transport of diatoms in the Antarctic Sirius Group: Pliocene refrigerator, *GSA Today*, *8*(4), 1–8.

Heimsath, A., J. Chappell, N. Spooner, and D. Questiaux (2002), Creeping soil, *Geology*, *30*(2), 111-114, doi:10.1130/0091-7613(2002)030<0111:CS>2.0.CO;2.

Hendy, C. H., T. R. Healy, E. M. Rayner, J. Shaw, and A. T. Wilson (1979), Late Pleistocene glacial chronology of the Taylor Valley, Antarctica, and the global climate, *Quaternary Research*, *11*(2), 172-184.

Hindmarsh, R. C. A., F. M. Van der Wateren, and A. L. L. M. Verbers (1998), Sublimation of ice through sediment in Beacon Valley, Antarctica, *Geografiska Annaler. Series A. Physical Geography*, *80A*(3-4), 209-219.

Intergovernmental Panel on Climate Change (2007), *Climate change 2007 : the physical science basis : summary for policymakers.*, IPCC Secretariat, Paris.

Kellogg, D. E., and T. B. Kellogg (1996), Diatoms in South Pole ice; implications for eolian contamination of Sirius Group deposits, *Geology*, *24*(2), 115-118.

Kennett, J. P., and D. A. Hodell (1995), Stability or instability of Antarctic ice sheets during the warm climates of the Pliocene, *GSA Today*, *5*(1), 1-10.

Klein, J., R. Giegengach, R. Middleton, P. Sharma, J. Underwood, Jr, and R. Weeks (1986), Revealing Histories of Exposure Using in situ produced 26Al and 10Be in Libyan Desert Glass, *Radiocarbon*, *28*(No. 2A), 547-555.

Kyle, P. R. (1990), McMurdo Volcanic Group-Western Ross Embayment: introduction, *Volcanoes of the Antarctic Plate and Southern Oceans. Antarctic Research Series. Am Geophys Union, Washington, DC*, 18–25.

Lal, D. (1991), Cosmic ray labeling of erosion surfaces - In situ nuclide production rates and erosion models, *Earth and Planetary Science Letters*, *104*, 424-439.

Lancaster, N. (2002), Flux of Eolian Sediment in the McMurdo Dry Valleys, Antarctica: A Preliminary Assessment, *Arctic, Antarctic, and Alpine Research*, *34*(3), 318-323.

Lewis, A. R., D. R. Marchant, A. C. Ashworth, S. R. Hemming, and M. L. Machlus (2007a), Major middle Miocene climate change and the extinction of tundra communities: Evidence from the Transantarctic Mountains, in *Antarctica: A Keystone in a Changing World—Online Proceedings of the 10th ISAES X, edited by AK Cooper and CR Raymond et al., USGS Open-File Report*, vol. 1047.

Lewis, A., D. Marchant, A. Ashworth, S. Hemming, and M. Machlus (2007b), Major middle Miocene global climate change: Evidence from East Antarctica and the Transantarctic Mountains, *Geol Soc Am Bull*, *119*(11-12), 1449-1461, doi:10.1130/0016-7606(2007)119[1449:MMMGCC]2.0.CO;2.

Lewis, A., D. Marchant, D. Kowalewski, S. Baldwin, and L. Webb (2006), The age and origin of the Labyrinth, western Dry Valleys, Antarctica: Evidence for extensive middle Miocene subglacial floods and freshwater discharge to the Southern Ocean, *Geology*, *34*(7), 513-516, doi:10.1130/G22145.1.

MacClune, K., A. Fountain, J. Kargel, and D. MacAyeal (2003), Glaciers of the McMurdo dry valleys: Terrestrial analog for Martian polar sublimation, *Journal of Geophysical Research*, *108*(E4), 5031, doi:10.1029/2002JE001878.

Marchant, D. R., A. R. Lewis, W. M. Phillips, E. J. Moore, R. A. Souchez, G. H. Denton, D. E. Sugden, N. Potter, and G. P. Landis (2002), Formation of patterned ground and sublimation till over Miocene glacier ice in Beacon Valley, southern Victoria Land, Antarctica, *Bulletin of the Geological Society of America*, *114*(6), 718-730.

Marchant, D., and G. Denton (1996), Miocene and Pliocene paleoclimate of the Dry Valleys region, Southern Victoria land: a geomorphological approach, *Marine Micropaleontology*, *27*(1-4), 253-271.

Marchant, D., G. Denton, D. Sugden, and C. Swisher (1993b), Miocene Glacial Stratigraphy and Landscape Evolution of the Western Asgard Range, Antarctica, *Geografiska Annaler. Series A, Physical Geography*, *75*(4, A Special Volume Arising from the Vega Symposium: The Case for a Stable East Antarctic Ice Sheet), 303-330.

Marchant, D., G. Denton, and C. Swisher (1993a), Miocene-Pliocene-Pleistocene Glacial History of Arena Valley, Quartermain Mountains, Antarctica, *Geografiska Annaler. Series A, Physical Geography*, *75*(4, A Special Volume

Arising from the Vega Symposium: The Case for a Stable East Antarctic Ice Sheet), 269-302.

Marchant, D., G. Denton, C. Swisher, and N. Potter (1996), Late Cenozoic Antarctic paleoclimate reconstructed from volcanic ashes in the dry valleys region of southern Victoria Land, *Geol Soc Am Bull*, *108*(2), 181-194.

Marchant, D., and J. Head III (2007), Antarctic dry valleys: Microclimate zonation, variable geomorphic processes, and implications for assessing climate change on Mars, *Icarus*, *192*(1), 187-222, doi:10.1016/j.icarus.2007.06.018.

Marchant, D., C. Swisher, D. Lux, D. West, and G. Denton (1993c), Pliocene Paleoclimate and East Antarctic Ice-Sheet History from Surficial Ash Deposits, *Science*, *260*(5108), 667-670, doi:10.1126/science.260.5108.667.

Matsuoka, N., and K. Moriwaki (1992), Frost Heave and Creep in the Sor Rondane Mountains, Antarctica, *Arctic and Alpine Research*, *24*(4), 271-280.

McKay, C., M. Mellon, and E. Friedmann (1998), Soil temperatures and stability of ice-cemented ground in the McMurdo Dry Valleys, Antarctica, *Antarctic Science*, *10*(01), 31-38.

McKelvey, B. C., M. J. Hambrey, D. M. Harwood, M. C. G. Mabin, P. N. Webb, and J. M. Whitehead (2004), The Pagodroma Group–a Cenozoic record of the East Antarctic ice sheet in the northern Prince Charles Mountains, *Antarctic Science*, *13*(04), 455-468.

Molnar, P., and P. England (1990), Late Cenozoic uplift of mountain ranges and global climate change: chicken or egg?, *Nature*, *346*(6279), 29-34.

Montgomery, D. R., G. Balco, and S. D. Willett (2001), Climate, tectonics, and the morphology of the Andes, *Geology*, *29*(7), 579-582.

Morgan, D., J. Putkonen, G. Balco, and J. Stone (2008), Colluvium erosion rates in the McMurdo Dry Valleys, Antarctica, *Eos Trans. AGU*, *89(53), Fall Meet. Suppl.*, Abstract C34B-05.

Morris, E. C., T. A. Mutch, H. E. Holt, and G. US (1972), *Atlas of Geologic Features in the Dry Valleys of South Victoria Land, Antarctica: Possible Analogs of Martian Surface Features*, US Dept. of the Interior, Geological Survey.

Naish, T., R. Powell, R. Levy, G. Wilson, R. Scherer, F. Talarico, L. Krissek, F. Niessen, M. Pompilio, and T. Wilson (2009), Obliquity-paced Pliocene West Antarctic ice sheet oscillations, *Nature*, *458*(7236), 322-328.

Nash, D. B. (1980), Forms of bluffs degraded for different lengths of time in Emmet County, Michigan, USA, *Earth Surface Processes and Landforms*, *5*, 331-345.

Ng, F., B. Hallet, R. Sletten, and J. Stone (2005), Fast-growing till over ancient ice in Beacon Valley, Antarctica, *Geology*, *33*(2), 121-124, doi:10.1130/G21064.1.

Nichols, K. K., P. R. Bierman, W. R. Foniri, A. R. Gillespie, M. Caffee, and R. Finkel (2006), Dates and rates of arid region geomorphic processes, *GSA Today*, *16*(8), 4-11.

Nichols, R. (1966), Geomorphology of Antarctica, *Antarctic Research Series*, 1-46.

Nishiizumi, K. (2004), Preparation of 26Al AMS standards, *Nuclear Instruments and Methods in Physics Research Section B: Beam Interactions with Materials and Atoms*, *223-224*, 388-392, doi:10.1016/j.nimb.2004.04.075.

Nishiizumi, K., M. Imamura, M. Caffee, J. Southon, R. Finkel, and J. McAninch (2007), Absolute calibration of 10Be AMS standards, *Nuclear Instruments and Methods in Physics Research Section B: Beam Interactions with Materials and Atoms*, *258*(2), 403-413, doi:10.1016/j.nimb.2007.01.297.

Oehm, B., and B. Hallet (2005), Rates of soil creep, worldwide: weak climatic controls and potential feedback, *Zeitchrift fur Geomorphologie*, *49*, 353–372.

Ollier, C. D. (1979), Evolutionary Geomorphology of Australia and Papua: New Guinea, *Transactions of the Institute of British Geographers*, 516-539.

Prentice, M., and A. Krusic (2005), Early Pliocene alpine glaciation in Antarctica: Terrestrial versus tidewater glaciers in Wright Valley, *Geografiska Annaler*, *87*(1), 87-109.

Putkonen, J., G. Balco, and D. Morgan (2008a), Slow regolith degradation without creep determined by cosmogenic nuclide measurements in Arena Valley, Antarctica, *Quaternary Research*, *69*(2), 242-249, doi:10.1016/j.yqres.2007.12.004.

Putkonen, J., J. Connolly, and T. Orloff (2008b), Landscape evolution degrades the geologic signature of past glaciations, *Geomorphology*, *97*(1-2), 208-217.

Putkonen, J., and M. O'Neal (2006), Degradation of unconsolidated Quaternary landforms in the western North America, *Geomorphology*, *75*(3-4), 408-419.

Putkonen, J., M. Rosales, N. Turpen, D. Morgan, G. Balco, and M. Donaldson (2007), Regolith transport in the Dry Valleys of Antarctica, in *Antarctica: A Keystone in a Changing World—Online Proceedings of the 10th ISAES X, edited by AK Cooper and CR Raymond et al., USGS Open-File Report*, vol. U.S. Geological Survey Open-File Report 2007-1047.

Putkonen, J., R. Sletten, and B. Hallet (2003), Atmosphere/ice energy exchange through thin debris cover in Beacon Valley, Antarctica, , 913-915.

Putkonen, J., and T. Swanson (2003), Accuracy of cosmogenic ages for moraines, *Quaternary Research*, *59*(2), 255-261.

Schäfer, J. M., H. Baur, G. H. Denton, S. Ivy-Ochs, D. R. Marchant, C. Schlüchter, and R. Wieler (2000), The oldest ice on Earth in Beacon Valley, Antarctica: new evidence from surface exposure dating, *Earth and Planetary Science Letters*, *179*(1), 91-99.

Schafer, J., S. Ivy-Ochs, R. Wieler, I. Leya, H. Baur, G. Denton, and C. Schluchter (1999), Cosmogenic noble gas studies in the oldest landscape on earth: surface exposure ages of the Dry Valleys, Antarctica, *Earth and Planetary Science Letters*, *167*(3-4), 215-226.

Schiller, M., W. Dickinson, R. G. Ditchburn, I. J. Graham, and A. Zondervan (2009), Atmospheric 10Be in an Antarctic soil: Implications for climate change, *Journal of Geophysical Research*, *114*(F1), doi:10.1029/2008JF001052.

Schwerdtfeger, W. (1984), Weather and climate of the Antarctic, *Developments in Atmospheric Science*, *15*.

Selby, M. J. (1971), Slopes and Their Development in an Ice-Free, Arid Area of Antarctica, *Geografiska Annaler. Series A, Physical Geography*, *53*(3/4), 235-245.

Sletten, R. S., B. Hallet, and R. C. Fletcher (2003), Resurfacing time of terrestrial surfaces by the formation and maturation of polygonal patterned ground, *Journal of Geophysical Research*, *108*(E4), 8044, doi:10.1029/2002JE001914.

Staiger, J., D. Marchant, J. Schaefer, P. Oberholzer, J. Johnson, A. Lewis, and K. Swanger (2006), Plio-Pleistocene history of Ferrar Glacier, Antarctica: Implications for climate and ice sheet stability, *Earth and Planetary Science Letters*, *243*(3-4), 489-503.

Stone, J. O., R. S. Sletten, and B. Hallet (2000), Old ice, going fast: Cosmogenic isotope measurements on ice beneath the floor of Beacon Valley, in *Antarctica:*

Eos (Transactions, American Geophysical Union), Fall Meeting Supplement, Abstract H52C–21.

Stone, J. (2000), Air pressure and cosmogenic isotope production, *Journal of Geophysical Research*, *105*, 23753-23760.

Stone, J. (2004), Extraction of Al and Be from quartz for isotopic analysis, [online] Available from: http://depts.washington.edu/cosmolab/chem.html

Stroeven, A. P., M. L. Prentice, and J. Kleman (1996), On marine microfossil transport and pathways in Antarctica during the late Neogene; evidence from the Sirius Group at Mount Fleming, *Geology*, *24*(8), 727-730.

Sugden, D., D. Marchant, and G. Denton (1993), The Case for a Stable East Antarctic Ice Sheet: The Background, *Geografiska Annaler. Series A, Physical Geography*, *75*(4, A Special Volume Arising from the Vega Symposium: The Case for a Stable East Antarctic Ice Sheet), 151-154.

Sugden, D., D. Marchant, N. Potter, R. Souchez, G. Denton, C. Swisher III, and J. Tison (1995), Preservation of Miocene glacier ice in East Antarctica, *Nature*, *376*(6539), 412-414, doi:10.1038/376412a0.

Summerfield, M. A., F. M. Stuart, H. A. P. Cockburn, D. E. Sugden, G. H. Denton, T. Dunai, and D. R. Marchant (1999), Long-term rates of denudation in the Dry Valleys, Transantarctic Mountains, southern Victoria Land, Antarctica based on in-situ-produced cosmogenic 21Ne, *Geomorphology*, *27*(1-2), 113-129.

Taylor, G. (1914), Physiography and glacial geology of east Antarctica, *The Geographical Journal*, *44*(4), 365-382, doi:doi:10.2307/1778593.

Tooth, S. (2008), Arid geomorphology: recent progress from an Earth System Science perspective, *Progress in Physical Geography*, *32*(1), 81.

Warnke, D. A., B. Marzo, and D. A. Hodell (1996), Major deglaciation of east Antarctica during the early Late Pliocene? Not likely from a marine perspective, *Marine Micropaleontology*, *27*(1), 237-252.

Washburn, A. L. (1980), *Geocryology : a survey of periglacial processes and environments*, Wiley, New York.

van der Wateren, D., and R. Hindmarsh (1995), Stabilists strike again, *Nature*, *376*(6539), 389-391, doi:10.1038/376389a0.

Webb, P. N., D. M. Harwood, B. C. McKelvey, J. H. Mercer, and L. D. Stott (1984), Cenozoic marine sedimentation and ice-volume variation on the East Antarctic craton, *Geology*, *12*(5), 287-291.

Whitten, D., and J. Brooks (1987), *Dictionary of Geology*, Viking Penguin Inc, New York.

Wilch, T., G. Denton, D. Lux, and W. McIntosh (1993), Limited Pliocene Glacier Extent and Surface Uplift in Middle Taylor Valley, Antarctica, *Geografiska Annaler. Series A, Physical Geography*, *75*(4, A Special Volume Arising from the Vega Symposium: The Case for a Stable East Antarctic Ice Sheet), 331-351.

VITA

Daniel J. Morgan

Daniel J. Morgan was born in Portland, Oregon in 1980 to Alfred Leonard Morgan and Susan Dolowitz Morgan. Daniel went to Beaverton High School, where he graduated as valedictorian in 1998. From there, Daniel went to Pomona College in Claremont, California, where he met his future wife, Torrey Katherine Shelton Morgan, and graduated in June of 2002 with a double-major in Geology and Mathematics. After spending a winter skiing in Steamboat Springs, Colorado, Daniel enrolled at the University of Washington to pursue his PhD in Geological Sciences from the Department of Earth and Space Sciences.