Holocene Thinning History of David Glacier, Antarctica

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ABSTRACT

The Antarctic Ice Sheet is a significant component of the Earth System, modulating Earth's sea level and climate. Present day and projected ice mass losses from Antarctica are of paramount concern to human populations in low-lying communities around the world. Ocean freshening from future ice discharge events also has the potential to destabilise global climate patterns. Over 40 years of satellite observations have tracked changes in ice mass, extent and thickness in Antarctica. However, ice sheets respond on timescales that range from annual to millennial, and a geologic perspective is needed to fully understand ice sheet response on timescales longer than a few decades. This research seeks to provide an improved understanding of Antarctica's future by constraining its past.

I focus on one of the largest outlet glaciers in Antarctica, the David Glacier/Drygalski Ice Tongue system which drains the East Antarctic Ice Sheet, dissects the Transantarctic Mountains and discharges into the Ross Sea. I seek to answer two questions; (1) what is the timing and nature of David Glacier thinning since the Last Glacial Maximum approximately 20,000 years ago, and (2) what physical processes were responsible for the observed thinning? I answer these questions by mapping the terrestrial and marine geomorphology along the former margins and seaward extension of David Glacier, and by using surface exposure dating of bedrock and glacial erratics to constrain the timing of glacier thinning. I then use a numerical flowline model to identify the processes that drove glacier thinning and retreat.

Surface exposure ages from bedrock and glacial erratics at field sites both upstream and downstream of the modern grounding line reveal that David Glacier thinned for two millennia during the mid-Holocene. Near the coast, this thinning occurred at ~ 6.5 kya at a rapid rate of up to 2 m/yr. Upstream from the grounding line, the thinning was more gradual but occurred simultaneously with thinning downstream. The timing of glacial thinning at David Glacier correlates with thinning events at other glaciers in the region and is consistent with offshore marine geological records.

To identify the mechanisms responsible for the observed thinning of David Glacier, I conduct numerical model sensitivity experiments along a 1,600 km flowline, extending from the ice sheet interior to the continental shelf edge in the western Ross Sea. Offshore, the glacier flowline follows the Drygalski Trough,

where it crosses numerous grounding zone wedges of various sizes. The flowline and prescribed ice shelf width is guided by the orientation and distribution of mega-scale glacial lineations as well as overall sea floor bathymetry. I explore the response of a stable, expanded David Glacier to the effects of increasing sub-ice shelf melt rates, and decreasing lateral buttressing which may have occurred as grounded ice in the Ross Sea migrated southward of the David Glacier. These forcings were also combined to explore potential feedbacks associated with Marine Ice Sheet Instability.

This modelling demonstrates that David Glacier likely underwent rapid thinning over a period of ~ 500 years as the grounding line retreated to a prominent sill at the mouth of David Fjord. After a period of ~ 5 ka of stability, a second period of grounding line retreat in the model leads to the glacier reaching its modern configuration. This simulated two-phase grounding line retreat compares well with onshore geologically constrained thinning events at two sites (Mt. Kring and Hughes Bluff), both in terms of timing and rates of past glacier thinning. This retreat pattern can be forced by either increased ice shelf melting or reduced buttressing, but when combined, lower melt rates and less lateral buttressing is required to match onshore geologic constraints.

Together, the findings in this thesis provide new data to constrain the past behaviour of a significant portion of the East Antarctic Ice Sheet and critical insights into the mechanisms that control ice sheet thinning and retreat. Incorporation of these constraints and improved understanding of the underlying mechanisms driving glacier thinning and grounding line retreat will ultimately improve continental scale ice sheet models which are used to project the future behaviour of the Antarctic Ice Sheet and its influence on global sea level.

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My infatuation with Antarctica has been heavily influenced from my time at The Ohio State University. It was in Columbus, Ohio that I learned about the history and logistics of scientific research in the Antarctic. Guided by Terry Wilson, Anne Grunow, David Elliot, Larry Krissek and Peter Webb, I applied my interests in pursuit of a Masters of Science degree. Long chats in the deep field with fellow graduate students heightened my awareness that we were doing something special. It was here in Columbus, where I first learned about the legacy and can-do attitude of a small group of Kiwis punching above their weight in Wellington, New Zealand.

Here in Wellington, I have found another family of folks with a deranged sickness for Antarctic science. The faculty and staff of the School of Geography, Earth and Environmental Sciences along with the Antarctic Research Centre provide world-class support to ensure success of its students. Countless chats with fellow students and researchers have contributed greatly to new found friendships and collaboration. With overwhelming and enduring support from supervisors Andrew Mackintosh and Kevin Norton, this experience has provided me the academic footing required to continue Antarctic scientific research.

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DEDICATION

To those before who made it all possible.

Chapter 1

Introduction

1.1 Antarctica's future framed

The Antarctic Ice Sheet (AIS) is rapidly changing. Over 40 years of satellite data reveal large areas of mass loss over the Antarctic Peninsula, Amundsen Sea sector of West Antarctica, and portions of coastal East Antarctica ((Rignot et al., 2019), Figure 1.1).

Most of the AIS drains to the Southern Ocean and is ringed by ice shelves which act as a ‘safety band’, effectively preventing rapid discharge of marine-based ice which is grounded ice resting on bedrock below sea level ((Fürst et al., 2016), Figure 1.2).

Presently, thermal erosion beneath ice shelves driven by a warming ocean presents the greatest threat to future ice sheet stability. Where incursions of warm water are observed by in situ measurements, the thinning of ice shelves and subsequent increase in ice discharge is producing a rapid ice surface thinning signal that reaches far inland (Dutrieux et al., 2014; Pritchard et al., 2009). This ‘dynamic thinning’ process is currently underway but uncertainties remain regarding the potential for this process to trigger widespread collapse over major sectors of the ice sheet (Pritchard et al., 2009, 2012; Gudmundsson et al., 2019). In this collapse scenario, ice-shelf thinning causes a speed up in seaward ice flow which results in a subsequent draw down of the upper ice surface. This, together with grounding line migration into deeper waters along a landward-sloping bed profile leads to a positive feedback of ice mass loss also known as the Marine Ice Sheet Instability (MISI) ((Weertman et al., 1974; Mercer, 1978; Thomas and Bentley, 1978; Schoof, 2007; Pattyn, 2018; Meredith et al., 2019), Figure 1.3).

While potentially modulated by self gravitational and isostatic rebound feedbacks, this process is shown in models to produce rapid and sustained global sea level rise (Gomez et al., 2010; Barletta et al., 2018; Whitehouse et al., 2019). Cur-

Figure 1.1: (A) Ice speed of the Antarctic Ice Sheet (B) Change in flow speed from the time period 2007/2008 to 2014/2015 (C) Basin names for sub-regions and ocean temperature at 310 m depth from the Southern Ocean State Estimate (SOSE) (D) Bed topography between 0 and 1,100 m depth, with SLE of each basin in centimeters of SLE. (E) Change in grounding line ice discharge, D , from 1979 to 2017 for the 18 major sub-regions in billions of tons per year, circle radius proportional to change. (F) Total change in mass of major basins color-coded from blue (gain) to red (loss) for 1979-2017 with circle radius proportional to the absolute mass balance. (From (Rignot et al., 2019))

Figure 1.2: Spatial distribution of maximum buttressing of the Antarctic ice shelves from Fürst et al. (2016). For the labelled ice-shelf regions, ‘Passive Ice Shelf’ (PIS) indices provide information about sector specific amount of ice shelf loss is possible without initiating rapid ice discharge. Notably, there is very little PIS in the Amundsen/Bellingshausen Sea sectors.

Figure 1.3: Schematic representation of Marine Ice Sheet Instability (MISI), from (Meredith et al., 2019).

rently, large portions of the AIS are susceptible to MISI and therefore, have the potential to severely impact human populations in low-lying coastal settlements. Further, ice mass loss acts to freshen the global ocean and impact circulation which is modelled to alter global weather patterns and temperature distribution.

Decades of ground and satellite-based monitoring programmes provide real-time detection of changes and enhanced understanding of the feedbacks driving change, yet large outlet glaciers of the AIS respond on timescales of centuries to millennia. Geological reconstructions of the AIS can thus bridge the time gap by extending the observational record to provide a longer-term record of past glacial fluctuations.

Sediment and ice cores provide clues to past climate and suggest that glacial-interglacial ice sheet fluctuations were caused by changes in orbital parameters (i.e. Milankovitch cycles) related to Earth's position relative to the sun and resulting changes in the global temperature distribution ((Lisiecki and Raymo, 2005; Naish et al., 2009; Jouzel et al., 2007; Fudge et al., 2013; Patterson et al., 2014; Brook and Buizert, 2018), Figure 1.4). Further, detailed synchronisation of Arctic and Antarctic ice cores suggest a mechanism in which changes in one pole may have significant impacts for the other (Stenni et al., 2011; Pedro et al., 2018; Brook and Buizert, 2018)

The unique palaeo-perspective provided by sediment and ice cores, with a wealth of near-continuous measurements of ancient climate data, do have limitations. These include large uncertainties related to age dating and that drill sites reflect an integrated signal of past ice sheet/climate fluctuations (Andrews et al., 1999; Weber et al., 2011; Bentley et al., 2014). Significant advances in analytical capability and greater spatial coverage continue to improve on these limitations. Yet still, large international logistical and analytical investments related to obtaining and processing these long-term records continue to limit sector specific detection of mass loss. Given the large variation in ice sheet behaviour observed by current monitoring programmes, geological reconstructions of specific sectors are warranted. Further, given the marine-based nature of the AIS, marine and terrestrial studies are needed to provide a complete and robust reconstruction of any specific ice sheet sector (Bentley et al., 2014; Jones et al., 2015; Jeong et al., 2018).

Decades of polar marine geological and geophysical investigations provide evidence for the timing and dominant processes related to AIS advance and retreat. Ship-borne multi-beam mapping provides a high-resolution view of the sea floor morphology which allows identification of suites of geomorphic features associated with various glacial processes and behaviour ((Dowdeswell et al., 2016), Figure 1.5).

Figure 1.4: a) The EPICA Dome C δD b) The Dome Fuji δO^{18} c) and d) The EPICA Dome C and Dome Fuji dust records, respectively. e) and f) The EPICA Dome C/Vostok CO_2 and CH_4 records, respectively. g) Benthic oxygen isotope stack. h) and i) Records of abrupt climate variability. The records show abrupt Northern Hemisphere transitions (yellow vertical lines). The Antarctic cooling response of the bipolar seesaw lags by around two centuries behind the abrupt Northern Hemisphere events. From (Brook and Buizert, 2018) and references therein.

Figure 1.5: Suite of submarine glacial geomorphic features from Dowdeswell et al. (2016). These associations, based on decades of polar marine geophysical and geological mapping, form the framework in which to interpret processes and determine glacial advance and retreat timings.

Collection of short to intermediate length sediment cores allows for enhanced understanding of the controlling processes responsible for a particular sea floor geomorphic feature, development of climate and palaeoceanographic proxies, and constraints on timing through analytical measurements of marine fossils found in sediments (Bentley et al., 2014; McKay et al., 2016; Hillenbrand et al., 2017). This integrated marine science approach has led to a significant advance in understanding past glacial fluctuations, yet is limited by uncertainties related to dating marine sediments and the inability to detect changes in glacier thickness and thus robustly estimate ice volume change through time.

Terrestrial geology and geomorphology has a rich history extending back to the heroic era of Antarctic exploration (Priestley, 1923). Glacial geologists working along the margins of the ice sheet have identified past periods of ice thickening through observation of high elevation glacial deposits (Péwé, 1960). Glacial drift deposits along the low-lying sectors of the ice sheet margin contain marine fossil material which via radiocarbon dating, provides a means of determining the timing of glacial advance and retreat (Denton et al., 1989; Bockheim et al., 1989; Stuiver et al., 1981; Orombelli et al., 1990; Baroni and Hall, 2004). This approach is limited by the age range of radiocarbon dating (approximately Last Glacial Maximum to present), uncertainties associated with dating marine fossils in Antarctica, and the overall inability to reconstruct the palaeo-ice sheet in high-elevation environments

or at high temporal resolution.

Starting in the late 1990s, surface exposure dating, which exploits the production of cosmogenic nuclides in rocky materials at the earths surface, revolutionised the field of terrestrial glacial geology (Lal, 1991; Balco, 2011). Early studies quickly realised that through multiple glacial cycles, glacial materials were recycled and carried a memory of prior exposure (Briner and Swanson, 1998; Sugden et al., 2005). This inherited signal resulted in many ‘complex exposure histories which ultimately failed to detect coherent, high-resolution palaeo-tracking of the ice sheet. Starting with Stone et al. (2003) and through careful sampling at select field sites, low-inheritance, high resolution signals of past thinning began to emerge. Taken together with improvements in analytical detection limits, numerous palaeo-thinning events for specific outlet glaciers has effectively, extended the observation record into deep time (e.g Stone et al., 2003; Bentley et al., 2006; Mackintosh et al., 2007; Johnson et al., 2008; Todd et al., 2010; Jones et al., 2015; Spector et al., 2017; Goehring et al., 2019; Small et al., 2019). This extension of the observational record allows inferences to be drawn about processes that were previously only observed on decadal timescales. For example, high-resolution palaeo-glacier thinning studies have shown sustained rapid thinning can persist for centuries after initiation ((Small et al., 2019), Figure 2.2). Further, surface exposure studies combined with glacier flow models enable improved understanding of the processes controlling glacier behaviour at catchment scale, which provides a longer-term perspective on theoretical models based on modern observations (Jones et al., 2015).

The growing spatial coverage of past ice sheet fluctuations provides additional geological constraints for continental ice sheet modelling studies aimed at predicting the future fate of the AIS and its role in global sea level rise (Briggs and Tarasov, 2013; Golledge et al., 2015)). Models which integrate physical processes identified in the modern observational record are combined with palaeoclimate proxy data to reconstruct the past AIS and its contribution to past global sea level rise. Models which accurately predict ongoing changes as well as past glacial fluctuations, constrained by geological reconstructions, provide a means of projecting the future behaviour of the AIS and importantly, its role in modulating global sea level (Meredith et al., 2019). Currently, future projections contain significant uncertainties arising from implementation of critical ice-ocean-atmosphere processes identified in the short observational record. Further understanding of these earth system interactions and feedbacks is needed to overcome large uncertainties in future projections. Further, there is a need for more complete ice sheet model evaluation using high-resolution onshore to offshore geologic reconstructions, and better attribution of ice sheet changes to human and natural causes (Meredith

Figure 1.6: Projected Global Mean Sea Level (GMSL) under Representative Concentration Pathway (RCP) scenarios RCP2.6, RCP4.5 and RCP8.5. Results are based on AR5 results for all components except the Antarctic contribution. From (Meredith et al., 2019)

et al., 2019) (Figure 1.6).

1.2 Research focus and motivating questions

Despite a growing body of research aimed at detecting past AIS fluctuations, large swaths of the Antarctic cryosphere remain unconstrained over the last glacial period. These include significant portions of the Transantarctic Mountains (TAM), Marie Byrd Land, Amundsen Sea, and various sectors of East Antarctica. In particular, David Glacier, one of the largest glaciers in the TAM, is constrained only by marine data or by inference from outside of its main drainage area (Shipp et al., 1999; Domack et al., 1999; Stuiver et al., 1981; Orombelli et al., 1990; Strasky et al., 2009). Given the paucity of ice proximal, high-resolution constraints, this research aims to answer two main questions:

1. What is the onshore timing and style of David Glacier thinning since the LGM?
2. What are the drivers and controls on the observed glacier thinning?

To answer question one, I conducted field work along the David Glacier where I collected glacial erratics and bedrock from field sites adjacent to fast flowing ice. Through careful collection, physical and chemical processing, I extract and measure cosmogenic isotopes which are used to determine surface exposure ages. These ages allow for the detailed tracking of the upper ice surface through time.

To answer question two, I use a glacier flowline model to perform sensitivity experiments along a 1,600 km flowline from the East Antarctic Ice Sheet interior to the shelf edge in the western Ross Sea, a domain encompassing the full extent of the expanded David Glacier. Given the results from surface exposure dating

and that the expanded David Glacier coalesced with grounded ice in the Ross Embayment, I investigate the sensitivity of the David Glacier to enhanced ocean temperatures and ice sheet-glacier decoupling.

1.3 Thesis Structure

This thesis is structured in a traditional format. The first chapter provides introduction, background, motivating research questions, and highlights the contributions by the author, the author's supervisors, and international collaborators.

Chapter two details the methods employed for the two aims of the research. Here I outline the utility and application of cosmogenic isotopes as it pertains to glacial reconstructions, and the physical framework for flowline modelling approach.

Chapter three is an analysis of the basin-wide terrestrial and marine geomorphology as well as long-term landscape evolution in the David Glacier area. Here I establish a geomorphological framework in which field sampling was undertaken and provide the context in which I interpret surface exposure ages and constrain flowline model boundary conditions.

Chapter four presents the thinning history of David Glacier using cosmogenic nuclides. Here I combine field and laboratory data to calculate surface exposure ages on a suite of samples from four sites along the David Glacier. Together, I determine an along flow reconstruction of the timing and style of thinning over a 150 km portion from the ice sheet interior to the mouth of the David Fjord. I conduct a data-model comparison with continental scale ice sheet model reconstructions, compare thinning rates from local, regional and Antarctic-wide palaeo thinning histories as well as draw inferences on the modern observational record of David Glacier.

Chapter five presents the results of the glacier flowline modelling. Here, I evaluate model runs under idealised scenarios in order to identify the processes responsible for the observed glacier thinning. I compare modelled thinning rates with those derived from geological data and draw conclusions regarding data-model match.

Chapter six synthesises the main results from Chapters 3-5. Further, I highlight outstanding and new questions as well as future research identified during field, lab and modelling work.

1.4 Contributions by author, supervisors, organisations and international collaborators

The following research thesis reflects a body of work I designed and carried out based on a concept originally envisioned by supervisor Andrew Mackintosh. I conducted data analysis and interpretation in the final thesis write-up with edits and supervision from Andrew Mackintosh, Kevin Norton and Stewart Jamieson. Field work in December 2017 was undertaken by myself, supervisor Andrew Mackintosh, PhD student Ross Whitmore and mountaineer Bia Bouchinas with field support provided by Antarctica New Zealand, Southern Lakes Helicopters and Kenn Borek Air. Flights to and from Antarctica were provided by New Zealand Defense Force and U.S. Air National Guard. I conducted all physical and chemical laboratory procedures under the supervision of Kevin Norton (SGEES/VUW) and Greg Balco (Berkeley Geochronology Center). Additional unpublished samples and field planning were offered by Carlo Baroni, Maria Cristina and Stefano Casale from the University of Pisa. I conducted all glacier modelling experiments under the supervision of Stewart Jamieson at Durham University. Additional modelling discussions and assistance were offered by Richard S. Jones and Pippa Whitehouse (Durham). Richard S. Jones provided initial field planning support as well as notes on chemical procedures and plotting tools through ice-tea.org. Fruitful discussion and sharing of initial results on the topic of onshore to offshore glacial reconstruction for the Northern Victoria land/Western Ross Sea area were facilitated by a series of SCAR workshops contributed to by Carlo Baroni (Pisa), Jae II Lee (KOPRI), Stewart Jamieson (Durham University), Chiara Sauli (IGS), Hyun Hee Rhee (Korea University), Ester Colizza (IGS) and myself. Access to unpublished bathymetric data was provided by Joohan Lee (KOPRI) and Jenny Black (GNS).

Chapter 2

Methodology: Cosmogenic Nuclides and Glacier Flowline modelling

Abstract This research has two components: 1) to constrain the timing, style and magnitude of thinning along the David Glacier since the Last Glacial Maximum and 2) to better understand the mechanisms that drove glacier thinning. To achieve my aim, I analyse bedrock and glacial deposits for cosmogenic nuclide concentrations and conduct sensitivity experiments using a glacial flowline model. In this chapter, I explain: 1) the underlying theory and application of cosmogenic nuclides for surface exposure dating and 2) the glacier flowline modelling approach.

2.1 Constraints on outlet glacier palaeo-thinning using Cosmogenic Nuclides

2.1.1 Why cosmogenic nuclides?

Cosmogenic nuclides have revolutionized the field of glacial geomorphology by providing a means of quantifying the timescale of landscape evolution (Balco, 2011). Understanding landscape evolution requires use of a geochronometer to place an inferred event/process in time. For this research, I use *in situ* produced cosmogenic nuclides to determine the precise duration of time since first exposure of glacial deposits to the cosmic ray flux. Once deposited and exposed to cosmic rays, glacial deposits or glacially modified bedrock surface begin to accumulate cosmogenic nuclides. In this section, I describe the theory associated with cosmogenic nuclides and the application for surface exposure dating in Antarctica. I finish with description of field and laboratory methods as well as how exposure ages are calculated in this study.

2.1.2 Cosmic ray flux at earth's surface

Cosmic rays, first studied by Victor Hess in 1912, are high-energy particles originating from galactic supernova explosions. Following their journey through space, cosmic rays interact with earth's atmosphere, surface and subsurface. Two reviews, Gosse and Phillips (2001); Dunai (2010) provide a common theoretical framework for the majority of surface exposure studies while Balco et al. (2008); Dunai and Stuart (2009); Marrero et al. (2016) provide an updated, standard calculation scheme based on the aforementioned theory. The following is an abbreviated explanation of the cosmic ray flux and its applications to Antarctic glacial reconstructions.

Prior to arrival at earth's upper atmosphere, the primary cosmic ray flux interacts with Earth's magnetic field. Ultimately, the major control on the cosmic ray flux is determined by the cutoff rigidity. The equation for the cutoff rigidity is given in equation 2.1 where p is the particle momentum, c is the speed of light and e is the particle charge. The vertical cutoff rigidity, the minimum particle energy needed to penetrate earth's magnetic field, is related to geomagnetic latitude and is very low in polar regions. South of 58° latitude, nearly all cosmic ray particles exceed the rigidity cutoff and remain energetic enough to interact with the surface of the earth.

$$R = pc/e \tag{2.1}$$

Upon entry to earth's upper atmosphere, high energy particles that exceed the binding energies of atomic nuclei, produce a spallation reaction and creation of cascade of secondary cosmic rays. These secondary cosmic rays are subsequently divided into three components: Electromagnetic, Mesonic and Nucleonic. Most importantly for terrestrial studies is the nucleonic component, which is the only component with energies high enough to produce a spallation reaction at the earth's surface. Once the secondary cascade reaches the Earth's surface, *in situ* terrestrial cosmogenic nuclides (TCN) (e.g. ^{10}Be) are produced by a spallation reaction, which is a high energy process where an incoming neutron collides with and disaggregates a target nucleus into several lighter particles (Figure 2.1). There are many cosmogenic nuclides that allow investigation of a wide range of processes and timescales at the Earth's surface. Table 2.1 is a subset of TCNs used in this study modified from Gosse and Phillips (2001); Dunai (2010).

Figure 2.1: A) Schematic for production of the cosmic ray flux and its interaction with Earth's magnetic field and upper atmosphere. B) Schematic showing secondary cosmic ray cascade through Earth's upper atmosphere and production of cosmogenic nuclides in surface materials (e.g. bedrock and glacial deposits). Photo from Evans Heights along the David Glacier.

Nuclide	Half-life	Primary target minerals	Advantages	Disadvantages
^3He	stable	Ol, Pyx, Hbl, Gnt	^3He has a low detection limit on a conventional mass spectrometer. Extremely long exposures can be investigated because it is stable.	Correction for radiogenic / nucleogenic / magmatic ^3He is necessary, particular on rocks with old crystallization ages. Greater opportunity for inheritance of stable cosmogenic ^3He if the mineral grains have ever been exposed previously.
^{21}Ne	stable	Qtz, Ol, Gnt, Plag?	Extremely long exposures can be investigated because it is stable. ^{21}Ne does not diffuse as rapidly through quartz as ^3He does. Higher production rate than ^{10}Be .	Correction for radiogenic / nucleogenic ^{21}Ne is necessary. Greater opportunity for inheritance of stable ^{21}Ne if the mineral grains have ever been exposed previously.
^{10}Be	1.36 ± 0.07 Ma	Qtz, Mgnt, Plag?	Simple, stoichiometric chemistry of target mineral. Quartz is resistant to chemical weathering and is almost ubiquitous. Mostly produced through spallation and muon reactions on O and Si. Not a gas, relatively immobile.	Atmospheric ^{10}Be is potentially a significant source of contamination. ^{10}Be has the lowest production rate in quartz of all TCNs so young surfaces (e.g. <1000 yr) require large samples or not possible.

Table 2.1: Nuclides used in this study, their half-lives, target mineral, advantages and disadvantages. Modified from Dunai (2010); Balco et al. (2008); Dugan et al. (2008); Nishiizumi et al. (2007); Gosse and Phillips (2001); Lederer and Shirley (1978)

2.1.3 Cosmogenic isotope production rates

The production of cosmogenic nuclides at the earth's surface is primarily controlled by spallation. The formula for the production rate from spallation is given in equation 2.2:

$$P_{s,m}(z) = S_r \sum_k S_{el,s} P_{m,k}(0) C_k \exp(-z/\Delta_{f,e}) \quad (2.2)$$

where $P_{m,k}(0)$ is the sea-level, high-latitude production rate of species m by spallation of element k (atoms $\text{g}^{-1} \text{a}^{-1}$), S_r is the topographic shielding factor (unitless), $S_{el,s}$ is the geographical scaling factor for spallation reactions (unitless), C_k is the concentration of the element k (atoms $\text{g}^{-1} \text{a}^{-1}$), z is the depth (cm) and $\Delta_{f,e}$ is the effective attenuation length for fast neutrons (g cm^{-2}). The production is summed to give the total spallation production rate.

Geographic scaling describes the cosmic ray flux modulation by earth's atmosphere and geomagnetic field. Many scaling frameworks exist to correct for latitude, elevation, atmospheric pressure and geomagnetic field changes. Table 2.2 provides a brief description of each scaling framework and references to the original study. For this study, the community standard LSDn framework, which accounts for a time-dependant magnetic field fluctuation and allows for a nuclide-dependent scaling, is preferred (Lifton et al., 2014; Marrero et al., 2016).

2.1.4 Surface Exposure Dating

The total concentration of a cosmogenic nuclide in any geological surface is a function of it's exposure and erosion history and can be used to calculate either minimum exposure ages or maximum erosion rates (Equation 2.3, Granger (2006)). A surface exposure history can be simple (i.e. steady-state or continuous exposure) or complex (i.e. multiple episodes of burial or erosion). For samples with a constant erosion rate, the steady-state concentration (N) at the surface is given by:

$$N = \frac{P}{\frac{1}{\tau} + \frac{\rho\epsilon}{\Delta}} \quad (2.3)$$

where P is the production rate, τ is the radioactive mean-life, ρ is rock density g cm^{-3} , Δ is the attenuation length and ϵ is the erosion rate (cm yr^{-1}).

Antarctic subaerial erosion rates are among the lowest on Earth and on the timescales considered in this thesis, are considered negligible (Marrero et al., 2018). Therefore, the simple exposure age (T_{exp}) for a surface sample, assuming no erosion is given by:

$$T_{exp} = \frac{-1}{\lambda} \ln\left(1 - \frac{N(0, T_{exp})\lambda}{\sum_i P_i(0)}\right) \quad (2.4)$$

Abbr.	Reference	Description
St	Lal (1991); Stone (2000)	Time-independent (constant production rate). Based on two different types of neutron reaction counting methods (photographic emulsions and neutron monitors). Scaling factor inputs are geographic latitude and atmospheric pressure, based on the Stone (2000) modification.
Lm	Lal (1991); Nishiizumi et al. (1989)	Time-dependent version of St based on time-variation in the dipole magnetic field intensity.
Li	Lifton et al. (2005, 2008)	Time-dependent model based on neutron monitor measurements and incorporating dipole and non-dipole magnetic field fluctuations and solar modulation. The scaling factor is based on actual atmospheric pressure, solar modulation, and a cutoff rigidity calculated using trajectory tracing.
Du	Dunai (2001, 2000)	Time-dependent model based on neutron monitor measurements and incorporating dipole and non-dipole magnetic field fluctuations. The scaling factor is based on an analytically calculated cutoff rigidity and atmospheric pressure. A long-term mean for solar modulation is used in this model.
De	Desilets et al. (2006)	Time-dependent model based on neutron monitor measurements and incorporating dipole and non-dipole magnetic field fluctuations. The scaling factor is based on a cutoff rigidity calculated using trajectory tracing and the actual atmospheric pressure.
LSD (Sf)	Lifton et al. (2014)	Time-dependent model based on equations from a nuclear physics model and incorporating dipole and non-dipole magnetic field fluctuations and solar modulation. The scaling factor is based on actual atmospheric pressure, solar modulation, and a cutoff rigidity calculated using trajectory tracing. The total flux (protons+neutrons) is used to scale all reactions.
LSDn (Sa)	Lifton et al. (2014)	Time-dependent model based on equations from a nuclear physics model and incorporating dipole and non-dipole magnetic field fluctuations and solar modulation. The scaling factor is based on actual atmospheric pressure, solar modulation, and a cutoff rigidity calculated using trajectory tracing. Nuclide-dependent scaling is implemented by incorporating cross-sections for the different reactions.

Table 2.2: Scaling methods with original reference and short description from (Marrero et al., 2016).

where λ is nuclide decay constant (if unstable nuclide) and N is the measured cosmogenic nuclide concentration.

In Antarctica, exposure ages of glacial deposits may contain significant geologic scatter due to reworking from multiple glacial cycles, often referred to as ‘inheritance’. Typically, simple exposure histories are represented by the youngest exposure ages and interpreted to reflect the most recent exposure with older populations reflecting an inherited nuclide inventory associated with a complex exposure history (Stone et al., 2003; Mackintosh et al., 2007; Jones et al., 2015; Spector et al., 2017). Given extremely low erosion rates, this complex history is attributed to the shielding, non-erosive nature associated with cold-based ice burial (Sugden and John, 1976; Atkins et al., 2002; Joy et al., 2014; Marrero et al., 2018). During episodes of cold-based ice cover, production of cosmogenic nuclides cease while decay of radiogenically unstable nuclides continue. It is common to exploit the relative decay rates of unstable nuclides to provide more insight to exposure histories and identify statistical outliers (Lal, 1991; Sugden et al., 2005; Granger, 2006; Bentley et al., 2006; Storey et al., 2010; Hein et al., 2014; Jones et al., 2016; Rhee et al., 2019).

Simple exposure history is assumed when data plotted on the two-nuclide plot of Lal (1991) fall in the ‘steady-state erosion island’. In the case of a simple, long-lived exposure history, erosion estimates require longer exposure times that can be significant on million year time scales (Di Nicola et al., 2012; Marrero et al., 2018). See Chapter 3 for more discussion as it pertains to long-term landscape evolution studies used in this research.

2.1.5 Field sampling

In this thesis, I focus on understanding the glacial fluctuations from the Last Glacial Maximum to present-day. In chapter 3, I discuss the geomorphic framework in which samples are collected for surface exposure dating. In chapter 4, I discuss the palaeo-thinning history for David Glacier based on surface exposure ages. For this glacial reconstruction aim, I have employed the ‘dipstick’ approach of Stone et al. (2003), where glacial erratics are collected in elevation transects and are interpreted to track changes in the upper ice surface. Many studies have used the ‘dipstick’ approach to constrain past thinning rates and a recent review of palaeo-thinning records allows for a robust comparison with other palaeo-thinning records, modern observed ice sheet thinning and modelled palaeo-thinning from ice sheet models (Small et al., 2019).

Within the David Glacier basin and Terra Nova Bay area, numerous studies have identified areas of past glacial erosion and deposition but do not continuously

record the LGM to present-day behaviour of the David Glacier (Steed and Drewry, 1982; Orombelli et al., 1990; Baroni et al., 2004; Baroni and Hall, 2004; Strasky et al., 2007; Di Nicola et al., 2009, 2012). The David Glacier basin has a rare set of outcrops that cover the modelled limit of the LGM ice surface along its length (Whitehouse et al., 2012; DeConto and Pollard, 2016). For this study, I target outcrops and nunataks flanking the David Glacier adjacent to zones of modern high surface velocity along the length of the glacier.

In order to reconstruct a continuous thinning history for each field site while attempting to minimize geological scatter associated with complex exposure histories, I preferentially sampled glacial erratics displaying: 1) clear signs of glacial transport indicated by striations, glacial shaping (bulleted, faceted) and/or non-local lithology and 2) perched stably on local topographic highs or benches with clear signs of glacial erosion (e.g. striations, crescentic gouges, rochs moutonnes). This ‘glacial dipstick’ method has proved successful in many studies of palaeo-thinning of Antarctic outlet glaciers (Figure 2.2 (Stone et al., 2003; Mackintosh et al., 2007; Jones et al., 2015; Spector et al., 2017; Small et al., 2019). Further, at each site, glacially molded bedrock was sampled to understand the timing of erosion, burial or longer term (Cenozoic) exposure history

As outlined above, there is a wide variety of cosmogenic nuclides that can aid in understanding surface processes. For this study, I primarily use the cosmogenic nuclide ^{10}Be to calculate surface exposure ages for each sample because its production rate is relatively well understood, chemical processing methods are well established and it can be used for studies encompassing the last few million years of earth’s history (Norton et al., 2008; Balco et al., 2008). ^{10}Be is readily produced in quartz therefore, in the field, I chose to target quartz-rich samples such as sandstones and granites. In some cases, quartz bearing rocks were absent or directly adjacent to pyroxene-rich mafic rocks which are suitable for surface exposure dating using the stable cosmogenic nuclide ^3He . At each sampling site, the chosen sample was marked to indicate the top of the sample and a standard set of site information (e.g. photos, unique sample number, geographic location, shielding, geomorphic setting) was collected to aid sample prioritisation for laboratory processing and exposure age calculation.

2.1.6 Laboratory methods

In the field, ~ 100 samples were collected for this study. Since it is cost and time prohibitive to process all collected samples, I chose to focus on two sites: Hughes Bluff and Mt. Kring. The two sites contain plentiful glacial erratics over the entire elevation range and cover ~ 150 km both upstream and downstream of the

Figure 2.2: A) Generalised map of mountains flanking a major outlet glacier. B) Example plot of elevation vs surface exposure age in idealised thinning scenario of C) cross flow profile and D) along flow profile schematic.

modern grounding line. From each site, a subset of samples were chosen based on elevation coverage and those samples collected in an optimal geomorphic context (e.g. striated and faceted clasts perched on topographic highs).

Once in the laboratory, samples are photographed, cut, crushed, sieved and purified prior to chemical processing. Guided by marks made in the field, an optimal thickness <5cm was cut from the top of the sample. Once cut with a Bianco saw, an actual thickness is measured with a digital caliper. Top-cut samples were then crushed in a jaw crusher, sieved to a 250-500 micron fraction and washed with water to remove fine material. Samples were then magnetically separated using the Frantz isodynamic magnetic separator. For sandstones, the relatively high quartz purity required minimal magnetic separation to remove quartz grains with inclusions. The interlocking framework and multi-mineral nature of volcanic material (e.g. granite, basalt, dolerite) required additional separation techniques including magnetic, heavy liquids and/or feldspar froth flotation separation. First, a hand magnet removed all highly magnetic minerals followed by low and high amperage magnetic separation. Separates requiring additional purification were then separated using froth flotation separation of PRIME lab. Prior to froth flotation, a separate was archived and not separated using the froth float method due to the presence of ‘old’ carbon in soap used in the process (Nichols and Goehring, 2019). This is important here in case future surface exposure studies wish to use *in situ* ^{14}C to determine the samples surface exposure history. Finally, in cases where froth flotation did not achieve a suitable purity, heavy liquids separation was undertaken. During physical processing, excess sample material were photographed to provide a record of processing, which may be potentially useful for future surface exposure dating or other geological investigations such as provenance studies (e.g. Palmer et al. (2012)).

Chemical processing of purified quartz followed a standard method of Norton et al. (2008) following von Blanckenburg et al. (1996) which includes further sample purification, addition of ^9Be carrier, sample digestion, isolation of Beryllium using column chemistry, calcine BeO and packing of targets for AMS measurements. An initial warm and agitated soak in 10% hydrochloric acid (HCl) removes oxides coating the grains which further purifies the sample prior to chemical processing. Several rounds of warm, agitated dilute hydrofluoric acid (HF) etches and removes meteoric ^{10}Be that coats the grains. Samples are dried, weighed on a high precision balance and a known amount of ^9Be carrier is added to the sample. The sample is then digested in aqua regia and several rounds of HF to further purify the sample. The sample is then passed through a series of ion-exchange columns to isolate Iron, Beryllium, Aluminum, Titanium and Sodium. The beryllium separate is then combined with ammonia to precipitate Beryllium which is then calcined

using a bunsen burner to produce beryllium oxide (BeO). BeO is then combined with niobium and packed in a target specially designed for PRIME lab accelerator mass spectrometer (AMS) measurements. A procedural blank is processed alongside all sample runs to quantify any Beryllium unintentionally introduced during the chemical processing.

Samples analysed for ^3He were processed differently according to best practices related to mafic rocks and pyroxene purification techniques (Bromley et al., 2014; Balter, 2018). In collaboration with Dr. Greg Balco at the Berkeley Geochronology Center (BGC), I extracted pyroxene from samples of Ferrar dolerite for ^3He analysis by crushing, sieving to isolate a 0.125-0.25 mm grain size fraction, cleaning in dilute HCl and HF, removal of highly magnetic (Fe-Ti oxides) and non-magnetic (plagioclase) fractions using a Frantz separator, and hand-picking of resulting pyroxene concentrates to remove remaining contaminants. ^3He analysis utilised the BGC ‘Ohio’ system, which is a MAP-215 sector field mass spectrometer with updated detector and counting electronics, coupled to a fully automated gas extraction and purification line. Helium is extracted from mineral separates in a laser ‘microfurnace’ by encapsulating 10-30 mg aliquots in Ta packets and heating the packets under vacuum using a 150w diode laser with coaxial optical pyrometer, allowing feedback control of the laser to achieve a prescribed pyrometer temperature. In most cases, all He was released in a 15-minute heating step at 1100°C, and a subsequent step at 1200°C was not distinguishable from total process blanks measured on empty Ta packets (total process blanks contained $22,000 \pm 9000$ atoms ^3He). In no case did the second heating step include more than 3% of total Helium. Released gases are then reacted with SAES getters and frozen to activated charcoal at 12.5 K, and He is released into the mass spectrometer at 33 K. Total He abundances are quantified by peak height comparison with variably sized aliquots of custom-mixed He gas standards, calibrated by direct measurement of ^3He and ^4He pressures using Baratron capacitance manometers, containing between 2.2×10^{-14} and 1.2×10^{-11} mol He. Reported total uncertainties on He concentrations include uncertainties derived from counting statistics, blank subtraction, and reproducibility of gas standard analyses. Analysis of the CRONUS-P pyroxene standard on this system during 2017-2019 yielded $4.99 \times 10^9 \pm 0.19 \times 10^9$ atoms g $^{-1}$ ^3He and $3.69 \times 10^{13} \pm 0.11 \times 10^{13}$ atoms g $^{-1}$ ^4He (mean and standard deviation of 19 measurements) (Blard et al., 2015). Pyroxene in Ferrar dolerite is known to contain a nonzero concentration of noncosmogenic ^3He , which has been estimated from the lowest concentrations observed in these studies, in shielded samples or samples expected to have very young exposure ages, to be less than 5-6 Matoms/g (Ackert, 2000; Ackert and Kurz, 2004; Margerison et al., 2005; Kaplan et al., 2017). In addition, linear regression of ^3He (Balco,

unpublished data) and ^{10}Be concentrations in a set of pyroxenes with Holocene exposure ages, described by Eaves et al. (2018) and Collins (2015), provides an estimate of 3.2 ± 1.1 Matoms/g. As this value is indistinguishable from the lowest ^3He concentration observed in the present study (2.9 ± 0.7 Matoms/g from sample EHW-05, collected near the present ice surface and therefore likely to have a negligible exposure age), we adopt it as the best available estimate of noncosmogenic ^3He in this lithology and subtract it from total measured ^3He concentrations to compute concentrations of cosmogenic ^3He (see Chapter 4, table 4.4).

Bedrock samples collected by Carlo Baroni are introduced here for the first time. These samples, collected over multiple field seasons, are used in some cases for the only available, high-elevation constraint on surface exposure history and landscape evolution. The field and laboratory methods used by Baroni and his team are described elsewhere (e.g. Oberholzer et al. (2003); Strasky et al. (2007); Oberholzer et al. (2008); Di Nicola et al. (2009); Strasky et al. (2009); Di Nicola et al. (2012)) and raw AMS data have been used to recalculate exposure ages within the CronusCalc calculator (Marrero et al., 2016; Jones et al., 2019).

2.1.7 Calculation of Exposure Ages

Measurements of $^9\text{Be}/^{10}\text{Be}$ ratios were made at PRIME lab AMS facilities and ETH-Zurich, and corrected for $^9\text{Be}/^{10}\text{Be}$ ratios measured in procedural blanks. I determined the ^{10}Be concentration using the known amount of ^9Be carrier added during chemical processing. Using the concentration of ^{10}Be , analytical uncertainties and sample specific metadata, I calculated exposure ages using the Jones et al. (2019) online calculator that uses a simplified version of the CronusCalc framework of Balco et al. (2008); Marrero et al. (2016). All ages are calculated using the LSDn scaling scheme of Lifton et al. (2014) and are reported with internal errors and external errors. For comparison between field sites, total error is reported.

2.2 Flowline modelling of David Glacier

2.2.1 Introduction

Numerical modelling of the Antarctic Ice Sheet (AIS) is useful for understanding the processes responsible for observed and/or reconstructed ice sheet fluctuations, aids hypothesis testing and allows for predictions of future sea rise driven by AIS mass loss. In this thesis, I use a flowline model to understand the controls on the thinning and retreat behaviour of David Glacier derived from geological data. In this section, I describe the suitability, theory and previous applications of the

model.

2.2.2 Model suitability

In this thesis, I use a flowline model, initially developed for tidewater glaciers in order to understand grounding line migration in a computationally efficient framework and to conduct sensitivity experiments aimed at understanding the role of sub-ice shelf melt rate and lateral buttressing (Vieli and Payne, 2005; Jamieson et al., 2012; Whitehouse et al., 2017). The model has subsequently been adapted for modern Greenland outlet glaciers as well as Arctic and Antarctic palaeo-ice streams (Nick et al., 2010; Jamieson et al., 2012; Enderlin et al., 2013; Clason et al., 2016; Whitehouse et al., 2017). Compared with 3-D models, this model is relatively simple yet has robust tracking of grounding line migration using an evolving grid mesh and agrees well with ‘boundary layer theory’ which predicts non-linear, bed-controlled grounding line instabilities associated with ice resting on a reverse (landward sloping) bed slope (Schoof, 2007; Pattyn et al., 2012). Given the imperfect knowledge of bed elevation data the 1-D model is well suited to model the David Glacier where a significant portion of the bed elevation is poorly constrained (Fretwell et al., 2013). Further, the simple model allows for carrying out large ensembles of model runs and sensitivity experiments.

2.2.3 Theory

The underlying theory and governing equations are described in Vieli and Payne (2005) and here I describe the model pertinent to my research questions and recent updates. In Chapter 5, I describe the implementation of the model to the unique geometry and boundary conditions of the David Glacier along with description of sensitivity experiments.

The modelling undertaken in this thesis focuses on a $\sim 1,600$ km long flowband covering a diverse set of flow regimes from the higher-elevation, low-velocity Dome C area to the lower-elevation, high-velocity outlet glacier (David Glacier), ice shelf (Drygalski Ice Tongue) and ice stream beyond where it converged with grounded ice in the Ross Sea. Ice flow is modelled as a mechanically coupled ice sheet, ice stream and ice shelf flow governed by fundamental equations below.

Ice sheet flow is governed by the shallow ice approximation which assumes ice flow only from internal ice deformation and horizontal velocity defined as:

$$u = C \left(\frac{\delta s}{\delta x} \right) h^{n+1} \quad (2.5)$$

where n is Glens flow law exponent, s is ice surface elevation and C is a constant

given by:

$$C = \frac{2A(\rho_i g)^n}{n + 2} \quad (2.6)$$

As ice reaches flotation ice is assumed to spread unidirectionally along the flowline, ice shelf flow is balanced by:

$$2 \frac{\delta}{\delta x} h v \frac{\delta u}{\delta x} = \rho_i g h \frac{\delta s}{\delta x} \quad (2.7)$$

where ρ_i is ice density, g is gravitational acceleration and v is the vertically average effective viscosity defined as:

$$v = A^{\frac{-1}{n}} \left[\left(\frac{\delta u}{\delta x} \right)^2 \right]^{(1-n)(2n)} \quad (2.8)$$

Where A is the temperature dependant rate factor and $\frac{\delta u}{\delta x}$ is the effective strain rate.

For an ice stream, equation 2.8 is modified by including a basal friction coefficient β which is linearly related to the horizontal velocity u and is given by:

$$2 \frac{\delta}{\delta x} h v \frac{\delta u}{\delta x} - \beta^2 u = \rho_i g h \frac{\delta s}{\delta x} \quad (2.9)$$

2.2.4 Previous Applications and Updates

Recent applications using the model include testing the impact of a width term to calculate lateral buttressing along a coupled ice stream-shelf and enhanced treatment of ice shelf dynamics. Nick et al. (2010) and Jamieson et al. (2012) modelled depth (H) and width (W) averaged ice flow (u) using the following equation:

$$2 \frac{\delta}{\delta x} h v \frac{\delta u}{\delta x} - \beta^2 u + \frac{H}{W} \left(\frac{5u}{2AW} \right)^{\frac{1}{n}} = \rho_i g h \frac{\delta s}{\delta x} \quad (2.10)$$

The inclusion of the width term allows modelling of changing offshore trough width which is common in palaeo-ice streams and outlet glaciers, and further supported by geomorphic interpretations using high resolution bathymetry data (Jamieson et al., 2012; Livingstone et al., 2012, 2016).

Whitehouse et al. (2017) improved the ice shelf dynamics to incorporate the treatment of large horizontal grounding line movements that are expected for the palaeo-David Glacier based on the long term presence of the Drygalski Ice Tongue and evidence of extensive sub-ice shelf conditions inferred from marine sedimentary cores (Domack et al., 1999; McKay et al., 2008). In the model, the ice shelf

geometry evolves with time and variations in ice shelf thickness and extent are fed into lateral drag calculations (Whitehouse et al., 2017). Improved treatment of ice shelf dynamics allows exploration of the grounding line sensitivity to ice shelf feedbacks such as reduced lateral buttressing. Importantly for this study, a reduction in lateral buttressing is expected as the expanded David Glacier and grounded ice in the Ross Sea decouple, which has been suggested from interpretation of submarine geomorphic features in the western Ross Sea (Shipp et al., 1999; Halberstadt et al., 2016).

Oceanic forcing is responsible for initiating a complex feedback between sub-ice shelf melt (e.g. driven by warm water incursions (Dutrieux et al., 2014)), ice shelf buttressing and grounding line migration (Joughin et al., 2012; Alley et al., 2015). By incorporating ice shelf dynamics, the model allows investigation of the impacts of ice shelf processes on grounding line retreat and onshore thinning (Whitehouse et al., 2017).

2.2.5 Summary

This chapter has described the methods employed for two aims of this thesis: 1) cosmogenic isotopes as a geochronometer in glacially modified landscapes and 2) flowline modelling as an aid for understanding controls glacial behaviour. Interpretation of cosmogenic isotopic data allows for surface exposure dating of glacial deposits and glacially modified bedrock to track the upper ice surface through time. Well established methods for field sampling and site selection in addition to new standard methods for calculating exposure ages allow for local to Antarctic-wide comparisons with other palaeo thinning records (Marrero et al., 2016; Jones et al., 2019; Small et al., 2019). Glacier flowline modelling enables investigation into the physical processes that were responsible for driving ice retreat, including changes in buttressing and the feedbacks associated with progressive and unstable retreat into sub-glacial troughs (marine ice sheet instability) (Vielé and Payne, 2005; Schoof, 2007; Pattyn, 2010). Importantly for this thesis, improved treatment of evolving ice stream width and ice shelf dynamics allows exploration of controls on grounding line migration that induce upstream thinning recorded by geologic data.

Chapter 3

Geomorphology and Landscape Evolution, David Glacier Basin

Abstract

Geographic isolation of the Antarctic continent and 34 million years of glacial fluctuations have led to a unique suite of landscapes. High elevation sites record long-lived land surface preservation and low erosion rates typical of the persistent, cold, hyper-arid climate while low elevation sites record deep glacial erosion and freshly exposed erosional surfaces. Along the David Glacier, a rare set of rocky outcrops hold clues to the long-term history of this landscape which is dominated by the presence of the deeply incised David Fjord - a >300km long glacial trough dissecting the Transantarctic Mountains. Detailed geomorphic mapping combined with quantitative estimates of long-term erosion rates inform surface exposure studies aimed at unravelling the most recent geologic history of ice behaviour. Field relations provide evidence of glacial overriding up to 700 metres above the modern ice surface and subsequent glacial thinning across a wide swath of the David Glacier basin. Overall, geomorphic characterisation of five terrestrial sites combined with marine geomorphic mapping provide the regional context in which this research is undertaken.

3.1 Introduction

The Antarctic landscape has formed through the action of tectonic, fluvial and glacial processes. The late-stage breakup of Gondwana played a critical role in the geographic isolation of the Antarctic continent and development of the Antarctic Circumpolar Current (Lawver and Gahagan, 2003; Stickley et al., 2004). A global decline in atmospheric $p\text{CO}_2$ combined with geographic isolation of Antarctica likely contributed to cooler temperatures and enhanced precipitation which ultimately led to the growth of coalescing local ice caps and initiation of the Antarctic Ice Sheet (AIS) approximately 34 million years ago (Ma) (Kennett, 1977; Zachos et al., 1996; Barrett, 1996; DeConto and Pollard, 2003). Generally, the AIS expanded and contracted numerous times since its inception with the first 20 Ma history dominated by warm-based ice (i.e. available meltwater at the bed) (Naish et al., 2001; Pekar and DeConto, 2006; Jamieson et al., 2010; Pollard and DeConto, 2020; Paxman et al., 2019). Ice domes nucleated at the center of East and West Antarctica and ice flowed radially towards the margins eroding spectacular glacial troughs, some of which dissect the Transantarctic Mountains (TAM) and extend to the continental shelf. In particular, large glacial troughs dissecting the TAM likely follow palaeo-fluvial networks which themselves follow major regional faults (Salvini and Storti, 1999; Stern et al., 2005; Huerta, 2007). For example, in the Byrd Glacier area, isostatic rebound associated with selective linear erosion of major glacial troughs accounts for 50% peak elevation which together has produced >5000m relief (Stern et al., 2005).

Ice sheet expansion into the marine realm occurred first between 23 and 18 Ma, culminating in a step-wise cooling and initial expansion of the AIS to the continental shelf (Shevenell et al., 2004; Williams and Handwerger, 2005; Jamieson and Sugden, 2007; Levy et al., 2019). Based on evidence from sedimentary and ice cores, the AIS fluctuations shifted during the mid-Pleistocene from an obliquity to eccentricity dominated pacing, highly influenced by sea level changes (Jouzel et al., 2007; Jamieson and Sugden, 2007; Patterson et al., 2014; Levy et al., 2019).

Since 14-18 Ma, it is argued that Antarctic landscape modification has been isolated to areas of continental scale radial and offshore trough axes, enhanced by melt water flow. This is important as the David Glacier basin is one of these continental scale radial troughs and shows modern evidence of abundant melt water drainage (Smith et al., 2009). Local evidence from geomorphic mapping and cosmogenic isotope analysis, suggests some high elevation sites (>700 metres above modern ice) have remained ice free for at least 6 Ma (Di Nicola et al., 2012; Oberholzer et al., 2008, 2003). Lower elevation sites show restricted glacial fluctuations during the mid-Pleistocene to Holocene but are geographically limited

to the coastal outcrops of Terra Nova Bay and interior outcrops in the Ricker Hills (Orombelli et al., 1990; Strasky et al., 2009; Rhee et al., 2019). The difference between high and low elevation sites highlights a ubiquitous feature of previously glaciated terranes: the topographically controlled selective erosional nature of the Antarctic Ice Sheet (Sugden and John, 1976; Sugden et al., 2005; Stern et al., 2005).

The focus of this dissertation is to track the David Glacier fluctuations since the Last Glacial Maximum (LGM), which occurred approximately 20,000 years ago. Given the focus of this study, field sites have been selected adjacent to zones of rapid ice flow in order to capture a clear signal of the most recent glacial behaviour. This approach seeks to minimise the influence of multiple glacial overriding events which create additional complexity in disentangling past glacial behaviour. Below, I describe the modern David Glacier and its regional terrestrial and marine geomorphology. Within this framework, I establish the detailed geomorphic setting of five sites visited in 2017 in which sampling was undertaken for surface exposure dating using cosmogenic isotopic analysis. Finally, in order to provide a long-term, basin-wide estimate of landscape evolution, I end this chapter highlighting sites with complex exposure histories and an analysis of long-term erosion rates.

3.2 Modern David Glacier Basin

The modern day behaviour of David Glacier and its seaward extension, the Drygalski Ice Tongue, is constrained by numerous studies using a wide variety of remote sensing techniques and historical mapping (Frezzotti and Mabin, 1994; Frezzotti, 1997; Rignot, 2002; Macayeal et al., 2008; Wuite et al., 2009; Stearns, 2011). Over a ~ 40 year period (1979-2017), the surface mass balance of David Glacier is positive 7.5 ± 0.4 Gt/a which is up to 5 Gt/a higher than previously reported (Rignot, 2002; Wuite et al., 2009; Stearns, 2011; Rignot et al., 2019). Calculated discharge from calving and basal melt over this period exceeds the surface mass balance over the four decade observational period, resulting in a -84 Gt/yr mass balance, translating to 0.2 mm sea level (Rignot et al., 2019). Using satellite laser altimetry, six areas are interpreted to contain active subglacial lake systems that fill and drain on a monthly to yearly basis (Smith et al., 2009). This fill and drain behaviour has been shown to increase the short-term velocity of the Byrd Glacier, another large TAM outlet glacier, by as much as 10% (Stearns et al., 2008; Rignot et al., 2019).

Estimates of basal melt from beneath the Drygalski Ice Tongue using various remote sensing methods range from 16-29 ma^{-1} with the highest melt rates at the grounding line and decreasing seaward (Frezzotti, 1993; Rignot, 2002; Wuite et al.,

2009). Basal melt derived from the grounding line provides abundant fresh water for an active ice-pump mechanism by which cold, fresh, less dense water rises, decreases in pressure reaching a supercooled state allowing localised ice accretion at the base of the Drygalski Ice Tongue (Lewis and Perkin, 1986; Frezzotti et al., 2000; Wuite et al., 2009). Modern oceanographic observations of water temperature, salinity and vertical water column structure near the Drygalski Ice Tongue edge provide additional understanding of ocean-ice shelf interactions, specifically the ice tongue’s role in the modifying currents from the southern Scott Coast and formation of high-salinity shelf water in the Terra Nova Bay Polynya (Stevens et al., 2017).

Modern surface velocity for the David Glacier is well constrained by satellite interferometric synthetic-aperture radar data (Rignot et al., 2011). The advance rate of the Drygalski Ice Tongue front, observed over a 100-year period based on historical mapping, aerial photos and satellite imagery, roughly matches surface velocity, implying that small-scale regular calving is not the primary mechanism for mass loss at the terminus (Wuite et al., 2009). Large calving events resulting from iceberg collisions or storm events may be the dominant control on terminus position (Priestley, 1923; Frezzotti and Mabin, 1994; Macayeal et al., 2008; Wuite et al., 2009; Rignot et al., 2011).

3.2.1 Terrestrial Geomorphology

The David Glacier basin exhibits two major landscape types: large troughs dissecting the Transantarctic Mountains and alpine glacial topography adjacent to the large troughs. Broadly, the David Glacier descends from the East Antarctic ice sheet interior and drains into the Western Ross Sea (Figure 3.1A). Confluent ice streams, one originating from Dome C and another from Talos Dome, drain through two subglacial troughs which dissect the Transantarctic Mountains and extend inland for more than 300 km (Fig. 3.1a-b) (Steed and Drewry, 1982; Fretwell et al., 2013).

Upstream of the modern grounding line, a major subglacial trough is characterised by enclosed sub-basins connected by an undulating network of smaller troughs. Broadly, this large glacial trough suggests the subglacial geomorphology has formed through selective linear erosion (Fig. 3.1d) (Sugden and John, 1976; Jamieson et al., 2014b). Approximately coincident with subglacial lake locations and in areas of ice flow convergence, surface flow stripes appear discontinuously and converge as the David Glacier descends from the East Antarctic ice sheet interior into the David fjord. These longitudinal surface structures (commonly called flow-stripes, flow-bands or flow-lines) are flow-parallel curvilinear features

that form from deformation of primary firn stratification or in zones of compressional, accelerating ice flow or at the confluence between ice flow units (Glasser and Gudmundsson, 2012). Flow-stripes are visible on satellite imagery and have been mapped in many studies around the Antarctic continent (Ely and Clark, 2016). Using the Landsat Image Mosaic of Antarctica (LIMA) data set, I mapped all visible flow stripes for the David Glacier basin. Flow stripes extend throughout the floating Drygalski Ice Tongue and correspond with the along flow and cross flow jagged sub ice shelf morphology, which are likely favourable pathways for buoyancy-driven ocean water flow (Tabacco et al., 2000; Bianchi et al., 2001; Stevens et al., 2017; Dow et al., 2018).

Figure 3.1: Geomorphic overview of David Glacier area. Onshore-offshore map showing modern surface velocity of Rignot et al. (2011) and bathymetry Arndt et al. (2013) with five field sites indicated with black triangles (Hughes Bluff (HB), D'Urville Wall (DW), Evans Heights (EH), Trio Nunatak (TN) and Mt. Kring (MK), overlain by mapped flow stripes and marine geomorphic features. Profile locations indicated with black dashed line. Inset shows surface velocity, bathymetry, the extent of David Glacier's drainage basin (black line) and location of local ice domes Dome C and Talos Dome. Lower panels show along flow cross section showing relationship between field sites and modern surface and bed topography (Bedmap2) (A-A') as well as surface and bed topographic cross section through Hughes Bluff and D'Urville Wall (B-B') and through Ricker Hills and Mt. Kring (C-C').

David Glacier’s Cauldron (cover photo), a surface expression of a dramatic ice cascade, is an area of low-relief surface depressions and ridges, converged flow stripes and intense crevassing (Swithinbank, 1988). Observed glacial seismicity beneath the Cauldron, near the grounding line, provides clues to the erosive nature at the grounding zone (Rignot, 2002; Danesi et al., 2007; Zoet et al., 2012). Over a one year period, Zoet et al. (2012) observed approximately 20,000 seismic events at this site consistent with stick-slip behaviour of debris-laden ice moving over a single obstacle of rough bedrock. This suggests the importance of stick-slip behaviour, potentially modulated by subglacial meltwater flow and tidal forces, acting as a mechanism for enhanced subglacial erosion along the David Glacier. Based on the erosional model for Antarctica and the local relief ($\sim 4\text{km}$), this behaviour has likely endured since the Eocene/Oligocene transition (Sugden and John, 1976; Stern et al., 2005). Importantly for this study, this erosive behaviour plucks and entrains fresh rock clasts without prior exposure that remain shielded from accumulating cosmogenic nuclides until deposited at downstream sites.

Perched above the David Glacier, alpine glacial topography exhibits classic landforms such as horns, ridges, and cirques with smaller alpine style-glaciers draining into the David Glacier (Fig. 3.1c) (Skinner and Ricker, 1968; Orombelli et al., 1990; Capponi et al., 1999; Baroni and Hall, 2004; Danesi et al., 2007). Erosional (areal scouring, rounded summits, rochs moutonnes and glacial striations) and depositional land forms (tills, moraines and perched glacial erratics) are present at rocky outcrop sites around the area up to 700 metres above the modern ice surface (Baroni and Hall, 2004).

3.2.2 Marine geomorphology

In addition to establishing a thinning chronology, another objective of this thesis is to understand the processes controlling the LGM to present behaviour of the David Glacier using a flow line model. Given that the David Glacier expanded into a large swath of the western Ross Sea, an analysis of the geomorphology of the sea floor is necessary to reconstruct ice extent since the LGM and is critical for determining certain boundary conditions for flow line modelling.

Numerous studies around Antarctica have used ship-borne high resolution multi-beam bathymetric data to interpret and reconstruct palaeo-ice sheet behaviour and extent (Shipp et al., 1999; Wellner et al., 2006; Livingstone et al., 2012; Anderson et al., 2014; Dowdeswell et al., 2016). Using published interpretations and following established analogs, unpublished bathymetric models and the Global Multi-Resolution Topographic dataset (geomapapp.org), I mapped all glacial geomorphic features in order to constrain the areal extent of the expanded

David Glacier in the Western Ross Sea (Shipp et al., 1999; Ryan et al., 2009; Anderson et al., 2014; Halberstadt et al., 2016; Dowdeswell et al., 2016; Halberstadt et al., 2016; Lee et al., 2017). These sea floor features constrain the extent of grounded ice and provide evidence for a staggered retreat (Grounding Zone Wedges, GZW), streaming ice flow (mega scale glacial lineations (MSGSLs)), presence of free melt-water flow (channelized features) and proximal ice shelf conditions (Iceberg plough marks). Furthermore, the trend of MSGSLs within the Drygalski Trough informs flow line orientation, and the location of GZW's are used to compare simulated ice retreat to the observed ice flow history (Chapter 5).

The Drygalski Ice Tongue, begins to float 40 km inland from the present-day coastline and extends 100 km into the western Ross Sea. The ice tongue is over 1,600 metres thick at the grounding line and thins to approximately 250 m at the terminus (Swithinbank (1988); Frezzotti et al. (2000) and Operation Ice Bridge 2011). Radar profiles reveal a rippled bottom morphology likely related to the complex interaction between the ice and ocean circulation (Bianchi et al., 2001; Stevens et al., 2017). The entirety of the David fjord is an overdeepened basin exceeding 1600 metres below sea level (mbsl) with its seaward extent defined by a bathymetric sill rising to ~ 500 mbsl near the mouth of the David fjord between Hughes Bluff and D'Urville Wall. The bathymetric sill descends seaward into the Drygalski Trough, the deepest sector of the Ross Sea (~ 1300 mbsl). The Drygalski trough and adjacent shoals seaward of Nansen Ice Shelf contain abundant, spectacular streamlined bedrock features that grade to mega-scale glacial lineations (MSGSLs), features that form beneath streaming ice within a deforming till (Clark, 1993; Anderson et al., 2014; Dowdeswell et al., 2016). The MSGSLs extend northward for over 200 km suggesting grounded, streaming ice occupied this trough during the LGM. At this time, grounded ice in the Drygalski trough appears to have terminated between the 74°S and 73.5°S latitude, where the MSGSLs end abruptly and the sea floor is covered by numerous iceberg plough marks (Shipp et al., 1999; Domack et al., 1999; Licht et al., 1999, 2005). Two prominent GZW's occur near the continental shelf break and near Coulman Island while at least three smaller GZW's occur south of Coulman Island (Bart et al., 2000; Lee, 2019). Based on seismic stratigraphy and marine sedimentary records, the GZW's near the continental shelf break and Coulman Island are interpreted to represent grounding events prior to the LGM and at the LGM, respectively yet precise ages are poorly constrained (Shipp et al., 1999; Bart et al., 2000, 2011). Smaller GZW's in the Terra Nova Bay region provide information about local ice retreat to the coast, inferred to occur during deglaciation (Brancolini et al., 1995; Anderson et al., 2002, 2014; Halberstadt et al., 2016). These GZW's, isolated along

the perched western flank of the Drygalski Trough, represent localised periods of grounding line stability during retreat.

3.3 Geomorphology of Field Sites

Guided by the modern surface morphology along the David Glacier, I used a combination of satellite imagery, aerial and ground photography to establish a refined set of sites suitable for exposure dating. As a result of this initial reconnaissance, eleven areas containing ice-free bedrock perched above the modern ice surface along major flow lines both upstream and downstream of the present day grounding line were identified and ranked. The highest priority sample sites are located downstream from the present-day grounding line, where the most recent and dramatic changes in ice thickness are expected to be preserved in the geological record (Denton et al., 1989; Bockheim et al., 1989; Jones et al., 2015). Upstream of the present-day grounding line, I focused on sites proximal to areas of high ice surface velocity. These sites are expected to record slightly older (but still LGM to present) changes in ice thickness. Together, the sites provide a robust spatial coverage and have high potential to record any changes in David Glacier surface elevation between the LGM and the present day. Field work was carried out during the 2017/2018 Antarctic summer field season. Working together with Antarctica New Zealand and Italian National Antarctic Programme (PNRA), we undertook a ~ 3 week campaign in the deep field, camping at our highest priority site, Hughes Bluff. Once in the field, all eleven sites were visited and five sites were chosen for detailed mapping and sampling.

Within the context of initial mapping by Baroni et al. (2004), field observations from five distinct sites along the David Glacier are outlined below. Here I detail Hughes Bluff and D’Urville Wall near the coastal glacial outlet, Evans Heights and Trio Nunatak West near the modern day grounding line and Mt. Kring located ~ 100 km upstream from the modern day grounding line to provide a wide range of geomorphic field relations along the length of the exposed rocky outcrops within the David Glacier basin. The observed field relations form a framework in which to interpret surface exposure data discussed in the Chapter 4.

During field work conducted in 2017, helicopter based aerial photo surveys were conducted to create a high resolution digital elevation model and orthomosaic. We employed a method using structure from motion photogrammetry (SfMP) developed by Vargo et al. (2017). The results include four detailed aerial image compilations (orthomosaics) and digital elevation models (with derived products: hillshade, slope, aspect) to aid geomorphic mapping. Combining high resolution GPS measurements from the helicopter flight path and GPS measurements taken

from sample sites, the average horizontal resolution for the SfMP DEM is 0.2 metre and an average vertical error of 1.44 metres. The aerial survey combined with field observations allow for comprehensive mapping and landscape analysis at each site. For mapping the D’Urville Wall site, a high resolution aerial photo (<https://earthexplorer.usgs.gov/>) was orthorectified in a geographic information system and overlain on an 8 metre tile from the Reference Elevation Model for Antarctica (Howat et al., 2019).

3.3.1 Hughes Bluff

Along the southern margin of David Glacier, Hughes Bluff rises 200 metres above the modern Drygalski Ice Tongue (Figure 3.2). The outcrop consists of granitic bedrock with undulating, rounded topography exhibiting spectacular erosional features such as rochers moutonnées, glacial striations and crescentic gouges. These geomorphic features are formed by thick, wet-based ice overriding the site and flowing parallel to the modern flow direction. The majority of the bedrock is blanketed by a variety of glacial erratics comprised of granite, sandstone and dolerite clasts in a patchy deposit lacking finer grained matrix material. In areas of small connected valleys, a ‘steamrolled’ till deposit is present, consistent with persistent snow cover or cold-based ice interaction (Atkins, 2013). Adjacent to these ‘cold till’ deposits are notable areas of snow banks and blue ice, which is typically indicative of wind-driven accumulation and ablation. These areas were avoided for our sampling but are noted here as examples of ongoing landscape modification by less erosive ice. A low lying, emergent outcrop near the modern shear margin of the David Glacier provides clues to help disentangle the most recent changes along this section of the glacier. Here, a small patch of rounded and striated bedrock is covered in glacial erratics. Multiple supraglacial deposits are observed near this low-lying outcrop. Overall, sampling was avoided in these areas but they are noted here as potential analogs for the patchy deposit blanketing the majority of the outcrop. Sampling along Hughes Bluff was restricted to local high-elevation rounded summits along a vertical transect with plentiful glacial striae and erratics (Figure 3.3).

Figure 3.2: Hughes Bluff Orthomosaic and 20 metre contours derived from SfMP and Glacial geomorphic map of Hughes Bluff with mapped units, significant geomorphic features and sample sites. Insets include the LIMA dataset (Bindschadler et al., 2008) and surface velocity of (Rignot et al., 2011).

Figure 3.3: (a) Oblique aerial photo of Hughes Bluff with approximate location of topographic profile (red), (b) North facing view from low bedrock knob showing perched glacial erratics and location of lowest emergent bedrock knob, (c) Moulded, striated and grooved nature of bedrock at lowest outcrop of Hughes Bluff, (d) Example of spectacular rockés moutonnées observed throughout Hughes Bluff, (e) Example of 'cold till' observed in multiple localities along small scale channel forms (outlined in red), (f) Highest elevation outcrop of Hughes Bluff showing rounded summit morphology with glacial striations, view to east. (g) Topographic profile from SfMP showing representative sampling sites (red squares, orange circles).

3.3.2 D’Urville Wall

Approximately 20 km from Hughes Bluff and forming the northern margin of the David fjord, the D’Urville Wall is a dramatic cliff rising ~ 500 m above the David Glacier (Figures 3.4 and 3.5). The cliff separates the outcrop into low and high elevation sites. The low lying site contains abundant landforms of glacial erosion and deposition, nearly identical to Hughes Bluff. The outcrop consists of two glacially moulded flow parallel linear ridges draped in a matrix-poor, erratic-rich deposit. Striae are oriented parallel to the modern ice flow directions. In contrast to Hughes Bluff, the lower D’Urville Wall is flanked by a large lateral moraine deposit consisting of at least three organized longitudinal ridges of large blocks. Closest to the ice margin and within a complex shear margin, fresh rocky materials are emerging from within the ice. No samples were collected from this modern moraine but samples from similar deposits have been collected along the lower Mawson Glacier and provide insights to the recent/modern depositional environment along the glacier. Between the glacially moulded ridges and the Wall is a drape of material inferred to be supraglacial rockfall. The deposit consists of poorly organized, angular fragments of local granitic material. Taken together, the lower D’Urville Wall, being perpendicularly across the glacier from Hughes Bluff, may serve as an interesting tie point while also shedding light on processes associated with the modern Drygalski Ice Tongue behaviour.

The upper D’Urville Wall consists of a low profile outcrop of frost shattered and fractured bedrock with few scattered erratics and weathering pits near the edge of the cliff. Rising ~ 200 m from the upper D’Urville Wall outcrop is the rounded summit of Mt. Neumayer where faint glacial striae and weathering pits have been observed by Baroni et al. (2004). Here, striae are oriented in a more NE direction implying that Mt. Neumayer was once covered by overriding ice. The slight change in striae direction may reflect complex flow around the high summit or unconfined ice flow away above the confining influence of the David fjord. Forming the eastern boundary of the D’Urville wall site, a portion of a local ice cap flows over the cliff as an ice fall and joins the Drygalski Ice Tongue.

Figure 3.4: D'Urville Wall vertical aerial photo (<https://earthexplorer.usgs.gov/>) overlain by 20 metre contours derived from Howat et al. (2019) and Glacial geomorphic map of D'Urville Wall with mapped units, significant geomorphic features and sample sites. Insets include the LIMA dataset (Bindschadler et al., 2008) and surface velocity of (Rignot et al., 2011).

Figure 3.5: (A) Oblique aerial photo of D'Urville Wall with approximate location of topographic profile (red), (B) South facing view of modern moraine complex, field party circled in red, (C) Lower D'Urville Wall showing rounded topography and scattered glacial erratics, view to east, (D) West facing view of Upper D'Urville Wall showing flat topography and fractured bedrock, Mt. Neumayer in background, (E) Bedrock weathering pit from Upper D'Urville Wall, ice axe for scale, (G) topographic profile from SEMP showing representative sampling sites (red squares, orange circles).

3.3.3 Evans Heights

Located near the modern-day grounding line and flanking a zone of high surface velocity draining Talos Dome is Evans Heights, a granitic outcrop extending ~ 300 metres above the local ice surface (Figs. 3.6 and 3.7). The outcrop has three main ridges separated by local glacier ice and bounded by wind scoops and ice falls as this portion of the David Glacier descends into the David fjord. The site offers rare, continuous bedrock exposure through a zone of weathered bedrock. Up to ~ 100 metres above the ice surface, the bedrock is fresh with clear striae parallel to modern ice flow, rounded summits and a patchy blanket of glacial erratics. >100 metres above the modern ice surface, the bedrock is increasingly weathered with faint linear grooves (possibly striae), rounded summits, fractured bedrock, scattered erratics and weathering pits some of which contain glacial erratics. Erratics are of mixed lithology comprised of granite, sandstone (in some cases fossil wood) and dolerite. Field sampling was focused on the middle ridge as it contained the most continuous and traversable outcrop from the lowest elevation to the upper ‘plateau’ morphology occurring above ~ 300 metres above local ice surface. Numerous portions of the upper ‘plateau’ outcrop are of low relief ($<10\text{m}$) and surrounded by blue ice. Glacier ice at this site surrounds and follows the topography of the exposed outcrop. As ice descends around Evans Heights, a notable valley-like morphology adjacent to the bedrock, typically referred to as a wind scoop, has ~ 100 metres relief.

Figure 3.7: (A) Oblique aerial photo of Evans Heights with approximate location of topographic profile (red), (B) East facing view of lower elevation sites with perched glacial erratics and striated bedrock, (C) East facing view of mid elevation sites with perched glacial erratics and moderately weathered striated bedrock (D) South facing view of upper Evans Heights showing deeply weathered bedrock with erratic filled weathering pits, (E) West facing view from highest elevation outcrop of Evans Heights showing low relief outcrop and steadily rising ice surface to the west, (F) Topographic profile from SFMP showing representative sampling sites (red squares, orange circles).

3.3.4 Western Trio Nunatak

Western Trio Nunatak is located ~ 40 km upstream from the modern day grounding line of David Glacier and is comprised of Ferrar Dolerite bedrock with scattered rafts of Beacon Sandstone and remnants of fossil forests (Figs. 3.8 and 3.9 (Capponi et al., 1999)). Focusing on a northward facing ridge line flanking a zone of high surface velocity draining from Dome C, the outcrop is composed primarily of fractured bedrock and hexagonal patterned ground. Importantly, rare rounded summits form prominent benches that contain fresh glacial striae and scattered glacial erratics up to ~ 200 metres above the local ice surface. Observed striae are parallel to modern ice flow and erratics are primarily sandstones and dolerite. Two bedrock samples, one sandstone and one dolerite, were collected above the highest evidence for glacial overriding.

The lower ~ 10 m of the sampling transect at west Trio Nunatak is draped by a thin glacial till connected to a large lateral moraine deposit along the northern margin of the nunatak. The moraine deposit is comprised of numerous linear ridges (parallel to ice flow) with abundant ponds forming on top of the moraine. At the time of field work, these ponds were frozen but are reported to melt during higher temperatures in the middle of the austral summer. South of the moraine deposit is a prominent supraglacial, elliptical shaped deposit flanked by a steep bedrock ridge. A short visit to the western most prominent ridge revealed a large wind scoop, vertical cliff faces and an extensive field of patterned ground and a general lack of glacially modified bedrock or glacial erratics. The southern portion of this nunatak was not visited as it is the site of a long standing biologic research field site investigating endolithic fungal and bacterial communities within the rafts of Beacon Sandstone (Zucconi et al., 2016). The eastern margin of this nunatak is comprised of supraglacial rock fall and steep bedrock.

Figure 3.8: Western Trio Nunatak orthomosaic and 20 metre contours derived from SfMP and Glacial geomorphic map with mapped units, significant geomorphic features and sample sites. Insets include the LIMA dataset (Bindschadler et al., 2008) and surface velocity of (Rignot et al., 2011).

Figure 3.9: (A) Oblique aerial photo of western Trio Nunatak with approximate location of topographic profile (red), (B) South facing view of lower elevation sites with perched glacial erratics on bedrock, (C) West facing view of scattered glacial erratics on bedrock, (D) East facing view of uppermost striated bedrock, ice axe parallel to striation direction, (E) North facing view from highest elevation outcrop of west Trio Nunatak showing site of bedrock samples TN-BR-01 and 02 plotted in Fig 3.12B, (F) Topographic profile from SfMP showing representative sampling sites (red squares, orange circles).

3.3.5 Mt. Kring

Mt. Kring, an isolated nunatak ~ 150 km from the coastline, is an impressive sequence of horizontally-bedded volcanic strata associated with Jurassic magmatism related to the sill emplaced Ferrar Dolerite (Figs. 3.10 and 3.11 (Capponi et al., 1999)). The nunatak rises nearly 300 metres above the local ice surface and is surrounded by glacial flow stripes and ice falls, presumably associated with flow around a complex and rugged subglacial topography. Here, the lower portion of the outcrop is draped by a broad till drift sheet comprised of striated and moulded sandstone, dolerite and basaltic erratics. Focusing on two prominent ridge lines, erratics resting on striated and fractured bedrock were sampled from regular intervals over ~ 200 vertical metres. The sampling was completed in two transects: one along a $\sim$ N-S trending ridge line and a second extending to an E-W trending ridge line. The N-S transect, along the southeastern edge of the nunatak, contained abundant erratics in a thin, patchy till draping exposed bedrock up to a local rounded summit at ~ 120 metres above local ice surface. A low saddle between the rounded summit and the E-W ridge line was comprised of a thick till deposit and marks the upper limit of the drift complex. In 2009, Carlo Baroni visited Mt. Kring and sampled two glacial erratics from this thick till deposit. Extending up from the saddle, the E-W ridge line is comprised of serrated edges with a narrow, traverseable route to higher elevations. Along the transect, bedrock became increasingly fragmented with an absence of erratics >200 meters above the modern ice surface, where a small patch of striated bedrock was sampled. In all cases, striae are parallel to modern ice flow. From 200 metres above the local ice surface, the ridge line became discontinuous and untraversable. The prominent peak at ~ 250 metres above the modern ice surface, visited and sampled by Carlo Baroni, shows no sign of glacial striations and no erratics are observed.

Figure 3.10: Mt. Kring orthomosaic and 20 metre contours derived from SfMP and Glacial geomorphic map with mapped units, significant geomorphic features and sample sites. Insets include the LIMA dataset (Bindschadler et al., 2008) and surface velocity of (Rignot et al., 2011).

Figure 3.11: (A) Oblique aerial photo of Mt. Kring with approximate location of topographic profile (red), (B) West facing view of lowest elevation site with perched glacial erratics on bedrock, (C) West facing view of scattered glacial till draping bedrock ridge, (D) East facing view of scattered high elevation sandstone erratics perched on fractured bedrock, (E) East facing view from highest elevation striated bedrock outcrop, ice axe parallel to striation direction, (F) Topographic profile from SfMP showing representative sampling sites (red squares, orange circles).

3.4 Regional Landscape Evolution

Erosion is an important geomorphic process that influences landscape evolution. In Antarctica, erosion rates are some of the lowest on Earth due to cold, hyper-arid conditions and therefore a reduction in chemical weathering, soil development and biotic interactions responsible for enhanced erosion in more temperate environments (Portenga and Bierman, 2011; Marrero et al., 2018; Carracedo et al., 2019). Dominated by physical weathering processes such as abrasion by wind blown sediment, granular disintegration, frost/salt cracking, and a variety of subglacial processes associated with intermittent ice cover (warm versus cold-based ice cover), Antarctic erosion rates are primarily controlled by sediment supply, lithology and glacial history (Marrero et al., 2018; Sugden et al., 2005). Importantly, 'selective linear erosion' inferred from numerous glacial troughs found around Antarctica acts as a long-term erosion agent isolated to the troughs since the inception of the ice sheet ~ 34 Ma (Sugden and John, 1976; Stern et al., 2005). Based on local relief and flexural uplift modelling, this selective linear erosion is responsible for up to 50% of the long-term uplift and has eroded the troughs at ~ 0.06 mm/yr since the Mid-Miocene (Stern et al., 2005; Paxman et al., 2017).

In order to accurately account for erosion in calculation of surface exposure ages, long-term erosion rates need to be quantified (Lal, 1991). Multiple studies have investigated long-term erosion rates of high elevation bedrock sites within or near the David Glacier basin (Di Nicola et al., 2012; Oberholzer et al., 2003; Strasky et al., 2009; Marrero et al., 2018). A large variation in calculated erosion rates is likely due to calculations on bedrock versus erratics and high versus low elevation. Taking values from high elevation sites, erosion rates are less than 0.05 mm/kyr (Di Nicola et al., 2012). Upon recalculation using an Antarctic-wide reanalysis, Marrero et al. (2018) determined slightly higher yet still very low erosion rates 0.1 ± 0.01 mm/kyr. Given the extremely low erosion rates and relatively young ages in this study, erosion is considered negligible and assumed to be zero when calculating exposure ages.

Three new bedrock exposure ages from two sites allow inferences to be drawn involving past non-erosive ice cover. Relevant metadata and raw AMS measurements for these samples can be found in Table 3.1 and 3.2. These localities lie less than 100m above the highest erratic or evidence of areal scouring. Two bedrock samples from Western Trio Nunatak, in place and in direct contact, were sampled in order to quantify differential erosion rates and explore burial history (TN-BR-01, Beacon Sandstone and TN-BR-02, Ferrar Dolerite). A third bedrock sample, MK-BR-03, a bedrock from the highest point on Mt. Kring (~ 250 metres above local ice surface) quantifies maximum erosion rates and potential for longer term

ice cover.

Sample name	Lat	Long	Elevation	Pressure flag	Thickness	Density	Shielding	Erosion	Year Collected
961222.02	-75.232505	162.546648	353	ant	1	2.65	0.98	0	2009
961222.03	-75.233072	162.549615	353	ant	5	2.65	0.98	0	2009
961223.02	-75.26517309	162.2794516	703	ant	2	2.65	0.98	0	2009
961223.03	-75.27027778	162.2477778	473	ant	3.5	2.65	0.98	0	2009
100113.01	-74.988	157.904	1892	ant	2	3.25	1	0	2009
JS2017-MK-16-01	-74.99059	157.92499	1720	ant	4	2.94	0.9993	0	2017
JS2017-MK-BR-02	-74.98921	157.91114	1797	ant	4.1	2.94	0.9997	0	2017
TN-BR-01	-75.471	159.60379	1335	ant	3.9	2.7	0.999448	0	2017
JS2017-TN-BR-02	-75.471	159.60379	1335	ant	1	2.94	0.999448	0	2017
DN2012-ARC-1	-73.6965	162.64567	2725	ant	2	2.7	1	0	2012

Table 3.1: Relevant metadata used for calculating Bedrock exposure ages.

Sample name	Nuclide	Target Mineral	Concentration	Uncertainty	Standard	Standard Concentration
961222.02	Be-10	quartz	6231370	127400	07KNSTD	
961222.03	Be-10	quartz	3699000	136590	07KNSTD	
961223.02	Be-10	quartz	5547570	116980	07KNSTD	
961223.03	Be-10	quartz	964380	29440	07KNSTD	
100113.01	He-3	pyroxene	494957809	9230608	CRONUS-P	5031000000
JS2017-MK-16-01	He-3	pyroxene	351800000	12740000	CRONUS-P	5031000000
JS2017-MK-BR-02	He-3	pyroxene	440800000	14990000	CRONUS-P	5031000000
TN-BR-01	Be-10	quartz	2920020	42517	07KNSTD	
JS2017-TN-BR-02	He-3	pyroxene	139800000	6122000	CRONUS-P	5031000000
DN2012-ARC-1	Be-10	quartz	993000000	3000000	KNSTD	
DN2012-ARC-1	Ne-21	quartz	828100000	32000000	CREU-1	349000000

Table 3.2: Nuclide concentration for Bedrock samples used in this study

Following the sample processing method outlined in Chapter 2, extracted and measured ^{10}Be and ^3He are used to calculate bedrock exposure ages (Balco et al., 2008). Calculated exposure ages and maximum erosion rates are listed in table 3.3 and 3.4. Exposure ages for the two samples from Western Trio Nunatak are 164 ± 12.8 kyr (^{10}Be) and 269 ± 46.1 kyr (^3He) for TN-BR-01 and 02 respectively (Fig. 3.12b). Trio Nunatak bedrock samples do not plot in the ‘steady-state erosional island’, which provides a clear signal of prior burial. When compared with the oldest, highest elevation bedrock within the area, field geomorphic evidence and cosmogenic nuclide concentrations, sites considered above the local LGM limit, are likely to have been covered by LGM ice but this ice did not significantly modify the landscape allowing preservation of previously exposed and covered bedrock.

Regionally, using erosion rates determined from rare, high elevation sites with multiple nuclides provides the most robust measure of long-term erosion rates. Bedrock sites in the David Glacier basin are $< 1\text{Ma}$ and show evidence for complex burial, whereas one site from the top of the Archembault Range (DN2012-ARG1) displays paired nuclide concentrations consistent with steady state ‘simple’ exposure history (Fig 3.13 and 3.12). In particular, high-elevation, multi-Ma exposure ages are highly sensitive to erosion rates and ‘erosion adjusted’ exposure ages can yield significantly older ages (Lal, 1991; Fink et al., 2006; Di Nicola et al., 2012; Marrero et al., 2018). Overall, there are few sites within the David Glacier basin

Sample name	Nuclide	Age (yr)	Interr (yr)	Exterr (yr)	Age (kyr)	Total Uncert. (kyr)	Relative Elevation (m)
961222-02	Be-10 (qtz)	889859	22911	69974	889.859	92.885	303
961222-03	Be-10 (qtz)	497478	20855	39310	497.478	60.165	303
961223-02	Be-10 (qtz)	600963	14781	43919	600.963	58.7	653
961223-03	Be-10 (qtz)	110069	3454	7517	110.069	10.971	423
100113-01	He-3 (px)	560861	10460	62575	560.861	73.035	275
JS2017-MK-16-01	He-3 (px)	481351	17432	55744	481.351	73.176	103
JS2017-MK-BR-02	He-3 (px)	566325	19259	65205	566.325	84.464	180
TN-BR-01	Be-10 (qtz)	164064	2490	10390	164.064	12.88	210
JS2017-TN-BR-02	He-3 (px)	258229	11308	30573	258.229	41.881	210
DN2012-ARC-1	Be-10 (qtz)	3350569	262113	575087	3350.569	837.2	625
DN2012-ARC-1	Ne-21 (qtz)	3601517	139172	316734	3601.517	455.906	625

Table 3.3: Calculated surface exposure ages (LSDn scaling) for all bedrock samples in this study.

Sample name	Nuclide	Erosion rate (g/cm ² /yr)	Erosion rate (m/Myr)	Internal Uncert. (m/Myr)	External uncert. (m/Myr)
MK01CB	He-3 (px)	0.000285	0.878	0.0164	0.0979
TN-BR-01	Be-10 (qtz)	0.000966	3.58	0.0564	0.235
JS2017-TN-BR-02	He-3 (px)	0.00062	2.11	0.0923	0.25

Table 3.4: Calculated maximum erosion rates for bedrock samples in this study.

that rise 600 metres above the local ice surface and of those sites with exposure age data, a complex exposure history is inferred based on comparison with sites that have remained ice free since the Miocene (Di Nicola et al., 2012). While most bedrock samples have only one measured nuclide, future studies aiming to understand long term erosion-burial signals should incorporate a second nuclide, preferably a stable/unstable nuclide pair, to provide more clues on the long-term erosion history.

The David Glacier basin represents a diverse, glacially modified landscape displaying evidence of selective linear erosion along large troughs dissecting the TAM, glacial overriding of high elevation nunataks, local sub-glacial erosion and deposition at low-elevation outcrops and marine-based glacial advance and retreat. The evolution of the David Glacier basin is intimately linked to the long-term fluctuations of the David Glacier.

Based on available evidence and analogs with other TAM glacier catchments, the David Glacier likely began to dissect the TAM coeval with the inception of the AIS ~ 34 Ma. Since this time, erosion has been focused along the glacial trough dissecting the TAM. Selective linear erosion has been the greatest agent of erosion and landscape evolution by excavating a gateway to the open ocean. Since the early-Miocene, the David Glacier has expanded into the western Ross Sea multiple times with seismic data providing evidence of at least 5 major erosional unconformities (Bart et al., 2000). Since ~ 5 -7 Ma, the area has not experienced high

elevation (>700 metres above the local ice surface) glacial overriding (Di Nicola et al., 2012). Additionally, Pleistocene glacial overriding did not exceed 500 meters above the local ice surface (Strasky et al., 2009).

Since the LGM, the David Glacier has thinned from its LGM configuration and glacial erosion and deposition has been restricted to lower-elevation regions. The David Glacier did not expand to the shelf edge in the western Ross Sea during the LGM (Licht et al., 1996; Shipp et al., 1999; Domack et al., 1999; Bart et al., 2000). During deglaciation, the expanded David Glacier underwent grounding line retreat and onshore thinning (Stuiver et al., 1981; Anderson et al., 2014). The present day state (Surface Mass Balance and grounding line location) of the David Glacier is relatively stable in the marine realm (Frezzotti, 1993; Macayeal et al., 2008; Rignot et al., 2019). Intense modern sub-glacial erosion is likely occurring near the modern grounding line and along high velocity zones while ice-marginal deposits continue to emerge from the ice (Baroni and Hall, 2004; Rignot et al., 2011; Zoet et al., 2012).

3.5 Conclusion

Geomorphology and landscape evolution of the David Glacier basin provides important context for surface exposure studies and glacier flowline modelling (Chapter 4 and Chapter 5). Particularly for surface exposure studies, the local geomorphic context serves as framework in which exposure ages are interpreted. A clear and ubiquitous observation within the David Glacier is the presence of a glacially modified landscape. At lower elevation sites, fresh bedrock formed by enhanced erosion under wet-based ice. At higher elevation sites, bedrock is weathered yet still retains evidence of past ice behaviour. In attempting to understand the most recent glacial fluctuation (i.e. LGM-present), samples chosen within the optimal geomorphic context are expected to reveal a clear signal of the most recent glacial behaviour. This is accomplished through careful site selection at the regional scale and interpretation of the local scale geomorphic context in which samples are collected. Geomorphic context allows for a realistic adaptation of a flowline model to the unique geometry of the David Glacier basin (Chapter 5). Mapped subglacial and submarine morphology constrain the model profile and provide evidence of periods of staggered retreat from the outer continental shelf. In this way, the geomorphology serves as the essential framework in which the rest of this research is undertaken. In the following chapters, I discuss the thinning chronology and controls of the David Glacier behaviour since the LGM.

Figure 3.12: Two nuclide plots for: (A) Archambault Ridge Di Nicola et al. (2012) an example which plots within the 'steady state exposure' envelope implying a simple (single) exposure history and B) Trio Nunatak showing samples plot outside of the 'steady state exposure' envelope implying a complex exposure history of burial and re-exposure. See Figure 3.9E for location of B. Sample and uncertainty ellipses vary depending on scaling scheme employed. St is red; Lm is green; LSDn is blue. Plots made using online calculator of Balco et al. (2008)

Figure 3.13: Regional bedrock exposure ages plotted against site elevation above modern ice surface from selected sites from David Glacier area, including Archambault Ridge site from Di Nicola et al. (2012).

Chapter 4

Thinning History of David Glacier

Abstract

Recent estimates of Antarctica's contribution to future sea level rise have varying trajectories depending on dominant mass loss processes and anthropogenic emission pathways. In order to reduce uncertainty, enhanced understanding of processes responsible for observed mass loss remains a chief objective of international research efforts. While understanding modern processes is useful, the observation period is not long enough to fully understand the processes that might lead to major ice sheet disintegration. To meet this aim, glacial geologic reconstructions provide a record of glacier change over centuries to millennia, thereby extending the observation record to encompass the full temporal range of ice sheet variability. Here I present the first palaeo-thinning record for David Glacier, the largest East Antarctic outlet glacier in Victoria Land. Palaeo-ice surface elevation is reconstructed using geomorphic evidence and ^{10}Be and ^3He surface exposure dating of bedrock and glacial erratics collected from the modern ice surface to the highest local peaks. Thinning profiles from two sites, Hughes Bluff on the Scott coast and Mt. Kring situated in the interior of the East Antarctic Ice Sheet, show that David Glacier experienced a period of rapid thinning of up to $\sim 2\text{m/yr}$ ($0.19\text{--}2.06\text{ m/yr}$, 2σ) during the mid-Holocene ($\sim 6.5\text{ kya}$). Regionally, this thinning event correlates with episodes of rapid ice surface lowering and retreat at nearby outlet glaciers along the Transantarctic Mountains, showing that outlet glaciers along the Scott Coast of the western Ross Sea experienced simultaneous thinning, 3 kyr after the southern retreat of grounded ice in the Ross Embayment.

Potential mechanisms for regional outlet glacier thinning and retreat include a previously unrecorded influx of marine heat, marine ice sheet instability, and/or dynamic effects associated with an early Holocene separation from the retreating

ice sheet/shelf. Such rapid glacier thinning events are currently not captured in continental or regional scale numerical modelling reconstructions for this period. This is of concern because these models are currently used to predict the future ice sheet contribution to sea level rise.

4.1 Introduction

The Earth experienced its last major period of climate warming between $\sim 20,000$ and 11,700 years ago. During this period known as the Last Glacial Termination, ice sheets retreated causing a global sea level rise of $\sim 130\text{m}$ (Clark et al., 2009; Conway et al., 1999; Denton et al., 2010). Ice sheet thinning and retreat was at times rapid, contributing to sea level rise at a rate similar to the most extreme projected ice sheet contributions in response to anthropogenic warming (DeConto and Pollard, 2016). For example, during Meltwater Pulse 1a, approximately 14,000 years ago, sea level increased at a rate of ~ 3 meters per century for 500 years (Deschamps et al., 2012). Understanding the processes that cause Antarctic Ice Sheet (AIS) mass loss over centuries to millennia will help fully understand and predict the present-day, ongoing response of ice sheets to climate change.

Geological reconstruction of the AIS through time provide critical insights and can lead to improved understanding of Antarctica's contribution to past, present, and future global sea level and climate change. Modern and historical observations of the AIS only extend back ~ 100 years, and therefore do not capture the full variability and behavior of the ice sheet which responds on timescales over one to two orders of magnitude longer. A geological perspective is required to understand how the ice sheet evolves over centennial to millennial time scales, which has potential to provide insight into rates of ice sheet change expected in the near future. Further, integration of the data constraining marine extent and terrestrial thickness of an ice mass through time provides a robust data set for evaluation of numerical ice sheet model simulations, which are used to predict future ice sheet contributions to sea level. A geologic perspective on ice sheet behavior is particularly useful as:

1. Quantitative satellite observations of ice sheet changes only extend back ~ 40 years (IMBIE, 2018; IPCC, 2013; Miles et al., 2013; Rignot et al., 2019). Decadal observations do not fully capture natural variability in ice sheet behavior and are too short to detect the potential impacts of anthropogenic climate change (Jacobs et al., 2011; Holland et al., 2019).
2. Data from the Gravity Recovery and Climate Experiment (GRACE) satellites offer a powerful measure of present-day ice sheet mass change but require a correction for millennial to centennial-scale ice sheet load history. This is because the gravity signal is strongly influenced by crustal uplift from glacio-isostatic adjustment (GIA) (Whitehouse et al., 2012; Whitehouse, 2018). Solutions using GRACE gravity data corrected using a precise GIA signal have the capability to provide a more accurate estimate of

present-day ice sheet mass balance at a regional scale (King et al., 2012; Simms et al., 2019).

3. Examples of past ice sheet changes provide evidence about the climatic processes and dynamic feedbacks by which ice sheets respond to environmental forcing. For example, the mass balance of large, marine-terminating outlets of Antarctica is affected by decadal to interannual changes in regional climate drivers such as the Southern Annular Mode and El Nio-Southern Oscillation (Assmann et al., 2019; Dutrieux et al., 2014; Miles et al., 2013; Walker and Gardner, 2018; Wille et al., 2019). However, we do not currently know if these short-term drivers were important over centennial to millennial timescales.

Reconstructions of marine terminating outlet glaciers provide an opportunity to constrain and understand the terrestrial and marine sectors of a glacier. At Mackay glacier, reconstructed post-LGM terrestrial ice surface thinning profiles correlate to periods of staggered retreat as documented by geomorphic features on the sea floor (Jones et al., 2015). Glacier modelling studies show the rate and magnitude of this thinning and retreat at ~ 6.5 kya was primarily related to local topographic control as the ice retreated into an over-deepened basin. However, further south along the Transantarctic Mountains, multiple glaciers show a near simultaneous rapid thinning over the early Holocene (8-9 kya) which, together with other records along the Scott Coast, suggests an intimate connection with retreating ground ice in the Ross Embayment (Todd et al., 2010; Anderson et al., 2014; Spector et al., 2017). These studies show the importance of a well-constrained thinning record linked with offshore chronologies to disentangle the potential forcings and feedbacks associated with ice sheet retreat.

Today, David Glacier drains the East Antarctic Ice Sheet (EAIS), traverses and incises the Transantarctic Mountains and discharges into the western Ross Sea as the floating Drygalski Ice Tongue, which was first discovered by the British National Antarctic Expedition (1901-1904). Comprising an area of roughly the size of Aotearoa New Zealand ($\sim 210,000$ km²), the glacier is the largest in Victoria Land, representing a significant element of the Antarctic cryosphere, draining from Dome C and Talos Dome. Based on marine geological and geophysical data, at the LGM, the David Glacier thickened and extended hundreds of kilometers into the western Ross Embayment (Anderson et al., 2014; Livingstone et al., 2012; Shipp et al., 1999). Here, I use surface exposure ages from erratic cobbles deposited on nunataks surrounding the David Glacier to reconstruct the terrestrial thinning history.

4.2 Methods

Surface exposure dating using *in situ* terrestrial cosmogenic nuclides has transformed the ability to reconstruct the ice sheet fluctuations in Antarctica (Stone et al., 2003; Mackintosh et al., 2007; Todd et al., 2010; Balco, 2011; White et al., 2011; Jones et al., 2015; Spector et al., 2017; Small et al., 2019). In this study, we use surface exposure ages of glacial erratics deposited on nunataks flanking the David Glacier, to reconstruct changes in ice surface elevation through time, and provide estimates of past outlet glacier thinning rates. Each sampling site then provides a critical tie point to compare with continental-scale ice sheet modelling experiments and modern ice sheet thinning rates derived from satellite observations.

4.2.1 Field Methods

For this thesis, field sampling focused on rocky outcrops and/or nunataks adjacent to the David Glacier both upstream and downstream of the present-day grounding line. Downstream sites are expected to show the most recent and dramatic change while sites upstream are expected to record smaller and longer term changes in ice thickness (Anderson et al., 2004; Bockheim et al., 1989). Using structure from motion photogrammetric technique, we constructed a high resolution digital elevation model of each field site allowing integration of glacial geological field surveys with the regional and local geomorphology (see Chapter 3) (Baroni et al., 2004; Vargo et al., 2017). Sampling focused on glacially molded bedrock surfaces glacial erratics from local glacial deposits along vertical transects perpendicular to the modern glacial flow direction. Faceted and striated glacial erratics are preferred as they indicate subglacial origins with minimal post-depositional erosion. The sampling method allows reconstruction of changes in the ice surface through time (Stone et al., 2003; Mackintosh et al., 2007; Jones et al., 2015; Small et al., 2019). We selected both quartz-rich rocks for ^{10}Be cosmogenic nuclide concentration and mafic rocks for ^3He noble gas content.

Of the 110 samples collected in the field, 31 samples were processed for ^{10}Be analysis and 11 samples for ^3He . Additionally, 9 unpublished ages from Carlo Baroni have been incorporated into this study. For ^{10}Be , quartz is the target mineral and is present in granites and sandstones. In this case, sandstones are preferred as they contain $\sim 90\%$ quartz and require significantly less physical processing. In order to cover the full elevation range of the outcrop, granites were collected in cases where sandstones were absent. In the field, samples isolated to local topographic highs and distant from areas of blue ice, snow drifts or local till deposits are preferred. While it is the focus of this study to use glacial erratics

to track glacier thinning, bedrock samples were collected to understand long-term exposure and erosion.

4.2.2 Laboratory Methods

Samples collected in the field were processed at the cosmogenic nuclide facilities at Te Herenga Waka Victoria University of Wellington. For ^{10}Be , we separated quartz by crushing and sieving to extract the 250-500-micron grain size fraction. Magnetic minerals were separated using a Frantz isodynamic magnetic separator. For granitic rocks, feldspars were removed by froth flotation. Multiple etchings in 1-2.5% Hydrofluoric acid further purified the quartz separates. Beryllium was extracted from quartz following an established method including addition of ^9Be carrier, enhanced quartz etching and digestion followed by ion exchange column chemistry and BeO precipitation (Norton et al., 2008). Targets of BeO were packed and sent to the PRIME lab for analysis using accelerated mass spectroscopy (AMS). For ^3He , we targeted pyroxene from the 125-250 micron grain size fraction. Pure pyroxene was obtained using an established HF etching method and $^3\text{He}/^4\text{He}$ ratios were measured at the Berkeley Geochronology Centre noble gas spectrometer following established methods (Bromley et al., 2014; Blard et al., 2015; Balter, 2018). Exposure ages are calculated by converting the AMS derived $^{10}\text{Be}/^9\text{Be}$ ratio to ^{10}Be concentration by subtracting a known amount of ^9Be carrier added during chemical processing (Table 4.2 and 4.3). Final exposure ages are calculated using ^{10}Be or ^3He concentration, field information (elevation, shielding, and sample thickness) and production rate scaling method (LSDn), using Cronus-Calc systematics (see Chapter 2) (Balco et al., 2008; Lifton et al., 2014; Marrero et al., 2016; Jones et al., 2019).

Lab Sample Name	Field Sample Name	Latitude	Longitude	Elevation	Thickness	Density	Shielding	Year Collected
JS1	HB_TE_01	-75.39209	162.22773	68.186	7.29	2.7	0.991766	2017
JS2	HB_TC_01A	-75.39177	162.21246	120.794	4.79	2.3	0.996274	2017
JS3	HB_TE_11_A1	-75.39258	162.22579	79.095	5.44	2.3	0.979524	2017
JS4	HB_TC_03_A1	-75.39202	162.21335	124.313	5.72	2.3	0.995218	2017
JS5	HB_TC_05_AB	-75.39302	162.21218	159	4.75	2.3	0.999509	2017
JS6	HB_TC_06_A1	-75.39232	162.20908	146.351	4.56	2.3	0.998517	2017
JS7	HB_TC_07_A1	-75.39351	162.20915	164.244	5.54	2.3	0.989264	2017
JS8	HB_TC_11_A1	-75.39568	162.20212	185.293	5.44	2.3	0.991891	2017
JS9	HB_TC_13A	-75.39594	162.19969	203.507	2.20	2.3	0.998238	2017
JS10	HB_TC_15	-75.39738	162.19742	214.765	5.51	2.3	0.998346	2017
JS11	HB_TC_16_A1B1	-75.39854	162.19665	229.295	5.45	2.3	0.997228	2017
JS21	MK_04	-75.39227362	162.2258802	80.297	5.11	2.3	0.971888	2017
JS22	MK_05A	-75.39236179	162.2253738	78.3	4.29	2.3	0.990428	2017
JS23	MK_07A1	-75.3953803	162.2263892	85.094	5.13	2.3	0.992489	2017
JS24	MK_08A1	-75.39529155	162.2227002	107.003	3.97	2.3	0.999365	2017
JS25	MK_09A1	-75.39422057	162.2217412	112.251	4.49	2.3	0.998981	2017
JS12	MK_11	-74.99327	157.92598	1650.119	4.24	2.3	0.998778	2017
JS13	MK_15	-74.99319	157.92642	1660.003	5.01	2.3	0.997352	2017
JS14	MK_19A	-74.99261	157.92602	1677.801	5.41	2.3	0.999345	2017
JS15	MK_21A	-74.99224	157.92631	1681.466	5.97	2.3	0.998465	2017
JS16	HB-BR-02	-74.99124	157.9256	1698.841	4.85	2.7	0.996965	2017
JS17	HB-TE-12	-74.99059	157.92499	1723.156	4.65	2.3	0.99933	2017
JS18	HB-TE-14	-74.98938	157.91284	1787.44	4.08	2.3	0.918411	2017
JS19	HB-TE-13	-74.98962	157.9171	1753.518	4.56	2.3	0.993606	2017
JS20	HB-TE-09	-74.98989	157.91876	1738.885	3.27	2.3	0.996929	2017
JS26	MK01A	-74.99414	157.92555	1643	5.15	2.3	0.998387	2017
JS27	MK06	-74.99261861	157.9260405	1677.801	4.61	2.3	0.997697	2017
JS28	MK10	-74.99070254	157.9251142	1718.747	4.86	2.3	0.995998	2017
JS29	MK14	-74.99053983	157.924575	1720.13	3.44	2.3	0.996484	2017
JS30	MK20	-74.98972208	157.9179974	1744.766	4.65	2.3	0.999087	2017
JS32	MK16A	-74.98955063	157.9152023	1764.792	2.19	2.3	0.929692	2017
961216.28	961216.28	-75.39813219	162.1978236	250	4.5	2.65	0.98	2009
961222.02	961222.02	-75.23250454	162.5466485	490	1	2.65	0.98	2009
961222.03	961222.03	-75.2330717	162.5496152	490	5	2.65	0.98	2009
961223.02	961223.02	-75.26517309	162.2794516	720	2	2.65	0.98	2009
961223.03	961223.03	-75.27027778	162.2477778	550	3.5	2.65	0.98	2009
Mt. Kring 01	Mt. Kring 01	-74.988	157.904	1892	2	3.25	1	2009
Mt. Kring 02	Mt. Kring 02	-74.99	157.922	1718	2	3.25	1	2009
Mt. Kring 03	Mt. Kring 03	-74.99	157.922	1725	2	3.25	1	2009
EHW-05	EHW-05	-75.09823	161.49885	433	0.999	2.94	1	2017
EHW-15	EHW-15	-75.09473	161.49687	561	0.999	2.94	1	2017
HB-TC-02	HB-TC-02	-75.39177	162.21246	117	4.2	2.94	0.996	2017
HB-TC-12	HB-TC-12	-75.39568	162.20212	187	0.999	2.94	1	2017
HB-TE-02	HB-TE-02	-75.39209	162.22773	64	7.3	2.94	0.992	2017
MK-BR-02	MK-BR-02	-74.98921	157.91114	1797	4.1	2.94	0.997	2017
MK-12	MK-12	-74.99059	157.92499	1720	4.4	2.94	0.993	2017
MK-13	MK-13	-74.99054	157.92456	1719	4.2	2.94	0.997	2017
MK-16-01	MK-16-01	-74.99059	157.92499	1720	4	2.94	0.999	2017
MK-18	MK-18	-74.98953	157.91666	1755	3.6	2.94	0.994	2017
TN-BR-02	TN-BR-02	-75.471	159.60379	1335	0.999	2.94	1	2017

Table 4.1: Relevant field metadata for all samples mentioned in this study.

ID	Lab Name	Field Name	10Be/9Be (10 ⁻¹⁵)	Uncert. (10 ⁻¹⁵)	% Uncert.	Max 9Be Current (nanoAmps)	Min 9Be Current (nanoAmps)	Avg. 9Be Current (nanoAmps)	Fraction of Standard Current	Total 10Be counts
201801222	JS1	HB_TE_01	73.73	2.66	3.6	11218	4541	8122	0.995	832
201801218	JS2	HB_TC_01A	84.63	2.69	3.2	10712	7021	8828	1.082	1034
201801217	JS3	HB_TE_11_A1	89.97	2.68	3	10873	2737	5918	0.725	1300
201801220	JS4	HB_TC_03_A1	114.22	3.4	3	11932	7477	10025	1.228	1171
201801225	JS5	HB_TC_05_AB	136.51	3.96	2.9	11541	7682	10070	1.234	1290
201801219	JS6	HB_TC_06_A1	125.31	3.68	2.9	11422	5907	9236	1.132	1199
201801224	JS7	HB_TC_07_A1	132.97	3.85	2.9	10858	4880	8604	1.054	1565
201801221	JS8	HB_TC_11_A1	124.04	3.71	3	11724	4711	7475	0.916	1184
201801227	JS9	HB_TC_13A	64.72	1.95	3	12798	4623	8439	1.034	1268
201801223	JS10	HB_TC_15	116.43	3.67	3.1	11402	4120	7602	0.931	2156
201801226	JS11	HB_TC_16_A1B1	134.28	4.51	3.4	12081	6356	9579	1.174	1299
201801228	JSB1	Blank1	2.18	0.5	22.9	10544	4639	8300	1.017	23
201801339	JS12	MK_04	1422.25	15.13	1.1	10278	6458	8446	1.035	12381
201801340	JS13	MK_05A	897.58	10.86	1.2	10334	7098	9596	1.176	8872
201801341	JS14	MK_07A1	3496.22	27.94	0.8	10747	6413	9174	1.124	33041
201801342	JS15	MK_08A1	342.22	7.11	2.1	12132	5844	9721	1.191	5638
201801343	JS16	MK_09A1	2138.74	21.52	1	8506	3354	6134	0.752	13780
201801344	JS17	MK_11	2558.27	22.34	0.9	10610	4064	7957	0.975	21156
201801345	JS18	MK_15	311.51	6.88	2.2	8158	4359	6698	0.821	2182
201801346	JS19	MK_19A	3120.5	30.68	1	8275	3264	5869	0.719	19098
201801347	JS20	MK_21A	435.91	7.19	1.6	10464	8863	9761	1.196	4331
201801348	JSB2	Blank2	3.43	0.63	18.3	8818	2311	4407	0.54	30
201802555	JS21	HB-BR-02	104	3.6	3.5	7544	5460	6858	1.161	898
201802556	JS22	HB-TE-12	104.13	3.58	3.4	6335	2988	4271	0.723	909
201802557	JS23	HB-TE-14	103.9	4.36	4.2	5553	3352	4527	0.766	597
201802558	JS24	HB-TE-13	120.75	3.58	3	6454	3118	4693	0.794	1246
201802559	JS25	HB-TE-09	116.88	3.86	3.3	7674	5090	6736	1.14	989
201802560	JS26	MK01A	349.64	6.99	2	8056	6134	7341	1.242	3210
201802561	JS27	MK06	2524.02	30.02	1.2	7654	6725	7306	1.236	17197
201802562	JS28	MK10	6410.77	81.12	1.3	7469	6443	6872	1.163	41145
201802563	JS29	MK14	1753.55	26.31	1.5	6277	2856	4643	0.786	7800
201802565	JS30	MK20	2917.79	66.75	2.3	7702	6617	7322	1.239	19832
201802566	JS32	MK16A	169.03	3.99	2.4	7762	3774	5947	1.006	2192
201802568	JSB3	Blank3	5.9	0.77	13.1	7682	4283	6364	1.077	59

Table 4.2: AMS measurements for all ¹⁰Be samples measured at PRIME labs.

Lab Name	Field Name	AMS Batch	Qtz mass (g)	Carrier added (g)	9Be (g)	9Be uncertainty (g)	10/9 ratio	10/9 ratio uncertainty	10Be conc. at/g	10Be conc. Uncert	% uncertainty
JS1	HB_TE_01	1	63.7	0.9922	0.000302621	5.81E-06	7.37E-14	2.66E-15	22717	969	3.61E+00
JS2	HB_TC_01A	1	47	0.9904	0.000302072	5.80E-06	8.46E-14	2.69E-15	35416	1368	3.18E+00
JS3	HB_TE_11_A1	1	68	0.9913	0.000302347	5.81E-06	9.00E-14	2.68E-15	26087	959	2.98E+00
JS4	HB_TC_03_A1	1	67.5	0.9916	0.000302438	5.81E-06	1.14E-13	3.40E-15	33548	1220	2.97E+00
JS5	HB_TC_05_AB	1	68.8	0.9934	0.000302987	5.82E-06	1.37E-13	3.96E-15	39535	1404	2.90E+00
JS6	HB_TC_06_A1	1	68	0.992	0.000302056	5.81E-06	1.25E-13	3.68E-15	36612	1316	2.94E+00
JS7	HB_TC_07_A1	1	69.8	0.9925	0.000302713	5.81E-06	1.33E-13	3.85E-15	37907	1345	2.89E+00
JS8	HB_TC_11_A1	1	63.9	0.9922	0.000302621	5.81E-06	1.24E-13	3.71E-15	38569	1403	2.99E+00
JS9	HB_TC_13A	1	31.2	0.9953	0.000303567	5.83E-06	6.47E-14	1.95E-15	40674	1538	3.02E+00
JS10	HB_TC_15	1	59.1	0.992	0.000302056	5.81E-06	1.16E-13	3.67E-15	39088	1478	3.15E+00
JS11	HB_TC_16_A1B1	1	65.3	0.9946	0.000303353	5.83E-06	1.34E-13	4.51E-15	41012	1619	3.36E+00
JS12	MK_04	2	51.9251	0.9947	0.000303384	5.83E-06	1.42E-12	1.51E-14	553937	12192	1.06E+00
JS13	MK_05A	2	30.74	1.0027	0.000305824	5.87E-06	8.98E-13	1.09E-14	594442	13548	1.21E+00
JS14	MK_07A1	2	46.1417	0.9956	0.000303658	5.83E-06	3.50E-12	2.79E-14	1535971	31980	7.99E-01
JS15	MK_08A1	2	48.1312	0.9955	0.000303628	5.83E-06	3.42E-13	7.11E-15	142810	4090	2.08E+00
JS16	MK_09A1	2	55.6535	0.993	0.000302865	5.82E-06	2.14E-12	2.15E-14	776489	16863	1.01E+00
JS17	MK_11	2	52.25	0.9939	0.00030314	5.82E-06	2.56E-12	2.23E-14	990463	20923	8.73E-01
JS18	MK_15	2	36.9328	0.9949	0.000303445	5.83E-06	3.12E-13	6.88E-15	169139	5018	2.21E+00
JS19	MK_19A	2	56.835	0.9955	0.000303628	5.83E-06	3.12E-12	3.07E-14	1112734	24032	9.83E-01
JS20	MK_21A	2	56.4908	0.9952	0.000303536	5.83E-06	4.36E-13	7.19E-15	155281	3968	1.65E+00
JS21	HB-BR-02	3	69.0106	0.9929	0.000302835	5.82E-06	1.04E-13	3.60E-15	28759	1228	3.46E+00
JS22	HB-TE-12	3	67.8807	0.9933	0.000302957	5.82E-06	1.04E-13	3.58E-15	29290	1244	3.44E+00
JS23	HB-TE-14	3	59.7741	0.9956	0.000303658	5.83E-06	1.04E-13	4.36E-15	33264	1647	4.19E+00
JS24	HB-TE-13	3	65.9535	0.9954	0.000303597	5.83E-06	1.21E-13	3.58E-15	35326	1333	2.97E+00
JS25	HB-TE-09	3	63.6745	0.9948	0.000303414	5.83E-06	1.17E-13	3.86E-15	35335	1442	3.30E+00
JS26	MK01A	3	57.2717	0.9953	0.000303567	5.83E-06	3.50E-13	6.99E-15	121746	3444	2.00E+00
JS27	MK06	3	63.2265	0.9944	0.000303292	5.82E-06	2.52E-12	3.00E-14	807153	18276	1.19E+00
JS28	MK10	3	61.3206	0.9952	0.000303536	5.83E-06	6.41E-12	8.11E-14	2118523	48765	1.27E+00
JS29	MK14	3	35.143	0.9952	0.000303536	5.83E-06	1.75E-12	2.63E-14	1008658	24666	1.50E+00
JS30	MK20	3	61.3042	0.9959	0.00030375	5.83E-06	2.92E-12	6.67E-14	964094	28854	2.29E+00
JS32	MK16A	3	20.135	0.9955	0.000303628	5.83E-06	1.69E-13	3.99E-15	164372	5244	2.36E+00

Table 4.3: Calculated ¹⁰Be concentrations for all samples analysed at PRIME lab.

Sample name	aliquot	He-4 concentration		Total He-3 concentration		Total He-3, weighted mean		Cosmogenic He-3	
		(Tatoms/g)	+/-	(Matoms/g)	+/-	(atoms/g)	+/-	(atoms/g)	+/-
EHW-05	a	108.6	1.8	3.76	1.67	2.91	0.70	-0.3	1.3
	b	108.0	1.9	4.43	0.87				
	c	107.7	1.9	1.60	0.75				
EHW-15	a	216.5	3.7	6.91	1.46	6.3	1.1	3.0	1.5
	b	179.9	3.1	4.43	1.40				
	c	178.3	3.1	7.73	1.52				
HB-TC-02	a	230.0	5.5	11.85	2.08	8.8	1.4	5.6	1.7
	b	195.9	3.4	8.16	1.73				
	c	178.9	3.1	7.49	1.67				
HB-TC-12	c	99.2	1.7	17.48	3.31	17.5	3.3	14.2	3.5
HB-TE-02	a	149.5	2.6	4.90	1.35	7.5	1.0	4.3	1.5
	b	150.4	2.6	9.25	1.26				
	c	152.3	2.6	8.11	1.40				
MK-BR-02	a	407.9	6.8	446.70	14.77	443.9	10.6	440.7	10.6
	b	432.1	7.3	441.07	15.12				
MK-12	a	245.3	4.2	86.39	4.41	85.8	3.0	82.5	3.1
	b	251.1	4.4	85.24	3.97				
MK-13	a	449.8	7.7	105.10	5.03	96.8	3.7	93.5	3.9
	b	440.2	7.6	86.81	5.50				
MK-16-01	a	536.1	9.3	353.31	12.42	354.2	7.4	351.0	7.5
	b	547.6	9.7	354.64	12.73				
	c	n.m.		354.82	13.34				
MK-18	a	206.1	3.6	39.10	2.92	43.9	1.6	40.7	2.0
	b	216.1	3.7	47.23	2.54				
	c	n.m.		44.47	3.17				
Mt Kring 01	a	171.44	0.91	502.55	6.80	483.22	8.36	479.98	8.43
	b	134.85	1.23	452.14	9.05				
	c	161.11	0.28	494.96	9.23				
Mt. Kring 02	a	318.14	0.55	31.60	1.79	37.94	2.40	34.70	2.62
	b	222.28	1.52	35.55	2.67				
	c	316.96	4.10	46.68	2.73				
Mt Kring 03	a	20.76	0.17	35.23	1.54	32.49	2.24	29.25	2.48
	b	43.19	0.07	35.87	3.47				
	c	50.46	0.09	38.31	1.74				
	d	25.28	0.18	25.87	2.24				
	e	21.78	0.04	27.17	2.21				
TN-BR-02	a	603.9	10.5	143.02	6.03	146.9	4.9	143.7	5.0
	b	n.m.		154.51	8.43				

Table 4.4: ^3He concentrations for all samples in this study

4.3 Results

4.3.1 Chronology

Of the 53 samples analysed in this study, 23 yield mid-Holocene exposure ages. Holocene aged samples are interpreted to have received minimal prior exposure (i.e inheritance) and suggest a simple, steady state exposure history. Focusing on two sites, I derive a high-resolution chronology of ice-surface changes over time for Mt. Kring and Hughes Bluff.

Mt. Kring, situated in the interior of the EAIS, lies along the flank of the ice stream draining Dome C. The nunatak, composed of layered dolerite, rises ~ 300 metres above the local ice surface. Of the 24 samples processed, three are dolerite bedrock and 21 erratics of mixed lithologies (Fig. 4.1 and Fig. 4.2) The surface at the peak of Mt. Kring lacks glacial striations or erratics and has a ^3He exposure age of 554 ± 91 kya. A discontinuous, steep ridge line between ~ 200 - 300 metres above local ice was not sampled due to inaccessibility. At ~ 200 metres above the local ice, the bedrock is striated parallel to modern ice flow direction and has a ^3He exposure age of 550 ± 80 kya. The highest elevation erratic is found at ~ 180 metres above the local ice with increasing erratic abundance culminating in a drift sheet covering nearly all bedrock below ~ 110 metres above local ice. Striated, bulleted cobbles and boulders of dolerite, basalt and sandstone are common in the thin, patchy drift covering the bedrock. The glacial drift is composite in age and we identify three populations of erratic exposure ages (Fig. 4.2b-c). The younger population spans from 7.4-5.5 kya (^{10}Be , $n=5$)(Fig. 4.2d). The older population spans from 51-25 kya (^{10}Be , $n=8$ and ^3He , $n=4$). The oldest, less defined erratic age population shows scattered evidence of glacial behavior between 123-63 kya (^{10}Be , $n=2$, ^3He , $n=2$). The older populations are likely an artifact of repeated burial by overriding ice and exposure (i.e. inheritance).

Figure 4.1: SfMP derived Orthomosaic map of Mt. Kring with all exposure ages and total errors listed. Erratics and bedrock ages plotted as circles and squares, respectively. ^{10}Be and ^3He derived ages are plotted in red and grey, respectively.

Figure 4.2: Calculated surface exposure ages for Mt. Kring. (A) All exposure ages. Erratics plotted as circles, bedrock as squares (^{10}Be , red and ^3He , grey). Error bars show total uncertainty. (B) Kernel density estimates for ^{10}Be data, (C) Exposure age vs. relative elevation for all ^{10}Be data and (D) Holocene data.

Hughes Bluff, situated on the Scott Coast, is a granite outcrop along the southern flank of the David Glacier. The outcrop is glacially scoured exhibiting spectacular rochers moutonnés, crescentic gouge marks and striations parallel to modern flow directions. Scattered glacial erratics blanket the entire outcrop. ^{10}Be exposure ages from 15 erratics span from 6.7–4.3 kya (Fig. 4.3 and 4.4a–b). The majority of glacial erratics are dated to between 6.7 and 6.2 kya (Fig. 4.4c–d). Two bedrock surface exposure ages from the highest and lowest outcrops ($20.97 \text{ ka} \pm 1.39$, 180 metres above local ice and $5.5 \pm 0.47 \text{ ka}$, 30 metres above local ice) suggest significant wet-based glacial erosion since the LGM. At 20 metres above local ice, similar bedrock and erratic ages provide evidence of recent emergence of this lower outcrop since 5.5 ka. Given the extensive glacial erosion, the LGM surface elevation at this site was likely considerably greater than 230m, the maximum elevation at Hughes Bluff. While the onset and magnitude of thinning prior to 6.7 kya is poorly constrained, the Hughes Bluff chronology indicate a period of rapid ice surface lowering during the mid-Holocene.

Figure 4.3: SfMP derived Orthomosaic map of Hughes Bluff with all exposure ages and total errors listed. Two bedrock ages plotted as squares and ages listed in italics.

Figure 4.4: Calculated surface exposure ages for Hughes Bluff. Plots made using ice-tea.org. Kernel density estimates for all data (A) and Holocene ages (C). Exposure age vs Relative elevation (metres above modern ice) for all data (B) and Holocene data (D)

In an effort to identify higher elevation glacial activity and long-term erosion history, field work was undertaken along the D’Urville Wall and Mt. Neumayer area. The D’Urville Wall is a steep, high-elevation (>400 m) granite outcrop delimiting the northern flank of David Glacier. The geomorphology and exposure ages from two field sites situated high above the David Glacier (Mt. Neumayer and Cape Phillipi) limit the reconstruction of the upper glacier surface during the LGM. Mt. Neumayer extending above the D’Urville Wall forms a rounded summit with striations sub-parallel to modern ice flow with few scattered erratics. In areas above 400m, bedrock samples contain weathering rinds and deep weathering pits filled with erratics (Baroni et al., 2004; Giorgetti and Baroni, 2007). Bedrock exposure ages from Mt. Neumayer (690 metres above local ice, 659 ± 74 kya), a terrace on top of the D’Urville Wall (500 metres above local ice, 119 ± 12 kya) and Cape Phillipi (300 metres above local ice, 541 ± 62 kya, 979 ± 120 kya) suggest either a thin cover of cold based ice or ice free conditions since the LGM. As discussed in Chapter 3, these high elevation bedrock samples are much younger

than exposure ages from bedrock at similar height above the local ice surface and suggests burial by non-erosive ice. Together with the geomorphic evidence from Hughes Bluff, bedrock exposure ages from the northern flank of David Glacier constrain the LGM upper surface between 300-690 metres above the local ice surface, a similar estimate from LGM age drift deposits from previous studies in Terra Nova Bay (Stuiver et al., 1981; Orombelli et al., 1990) (Fig. 4.5).

Figure 4.5: Cross section from Hughes Bluff to Mt. Neumayer combining the high-resolution chronology (red circles) from Hughes Bluff with unpublished ^{10}Be derived exposure ages from high-elevation bedrock samples sampled by Carlo Baroni (red squares). Bedrock ages older than Holocene are interpreted to represent bedrock with minimal ice cover or erosion and are used as constraints on upper ice surface elevation since the LGM. Elevation data source (Howat et al., 2019), bathymetry data source (Fretwell et al., 2013).

Two samples from Evans Heights, a site situated near modern David Glacier grounding line, yield Holocene ages yet with high uncertainties (Fig. 3.6). This site is important as local ice is derived from Talos Dome and thus may provide clues to relative timing of thinning for ice streaming from Dome C versus Talos Dome. The ^3He exposure ages from two Ferrar Dolerite erratics suggests this site experienced mid-Holocene glacier thinning, coincident with Hughes Bluff and Mt. Kring. Further, the lower sample is one of the lowest ^3He concentrations ever measured for a Ferrar Dolerite (Balco, pers. comm.). The initial constraints for Evans Heights provide a low-resolution, high-uncertainty glacier thinning record but future work using ^{10}Be from sandstone erratics may provide a coherent, low-inheritance, high-resolution thinning record for ice derived from Talos Dome.

Figure 4.6: Calculated exposure ages for Evans Heights. ^3He AMS measurements for five aliquots from two Ferrar Dolerite erratics. Weighted mean (WM, grey circle with red outline) reveals Holocene ages consistent with high-resolution chronologies from Hughes Bluff and Mt. Kring.

4.4 Discussion

High-resolution terrestrially derived history of David Glacier place modern observations in a long-term perspective and allows for local, regional and continent wide comparisons with other glacial histories and modelled ice sheet reconstructions. Locally, an onshore-offshore reconstruction is now more robust by incorporating the existing and emerging terrestrial and marine derived chronology by Jones et al. (2015); Hall (2009); Rhee et al. (2019); Anderson et al. (2014). This study allows comparison with palaeo-thinning records from other outlet glaciers across Antarctica (Small et al., 2019). Further, the palaeo-thinning record from David Glacier allows for a comparison with modelled ice sheet reconstructions.

High-resolution, low-inheritance exposure ages obtained from Hughes Bluff and Mt. Kring overlap in time yet reveal very different thinning styles with Hughes Bluff reflecting rapid, ice surface lowering and Mt. Kring reflecting a much slower, linear thinning signal. Further, the chronology from Mt. Kring implies glacier thinning reached far inland along zones of ice streaming. This critical site provides rare clues to ice behaviour from the high-elevation margins of the EAIS.

4.4.1 palaeo-Thinning Rates and Data-Model Comparison

Using the high-resolution chronology from Hughes Bluff and Mt. Kring, I derive a robust estimate of past thinning rates along David Glacier using a weighted least squares regression scheme within the iceTEA plotting tools (www.ice-tea.org). These reconstructed glacier thinning rates can be compared to modern thinning

rates derived from satellite data and continent scale ice sheet models (Jones et al., 2019). At Mt. Kring, I reconstruct a monotonic thinning event from 7.5-5.5 kya of up to 0.19 m/yr (0.06-0.19, 2σ) (Fig. 4.7a). At Hughes Bluff, I reconstruct a period of rapid thinning from 6.7-6.2 kyr of up to 2 m/yr (0.19-2.06 m/yr, 2σ) followed by ~ 4 kyr period of minimal thinning (Fig. 4.3b). The reconstructed palaeo-thinning along the David Glacier during the mid-Holocene is synchronous with the majority of Antarctic palaeo-thinning records (Small et al., 2019). Hughes Bluff represents one of the highest thinning rates observed in Antarctica, comparable to a reconstruction from nearby Mackay Glacier (Small et al., 2019; Jones et al., 2015).

Figure 4.7: Linear thinning rates estimated for mid-Holocene records using ice-tea.org from (Jones et al., 2019) for: a) Mt. Kring Holocene samples, b) Hughes Bluff samples with outliers identified and omitted using outlier detection in ice-tea.org.

The variation in thinning rates between Mt. Kring and Hughes Bluff is interpreted here to represent a record of past dynamic thinning, a process involving ice loss due to accelerated flow, ice-shelf basal melting and grounding line retreat (Pritchard et al., 2009, 2012). Dynamic thinning is well documented along modern marine terminating margins of the Greenland and Antarctic ice sheets yet identi-

fication in the geologic record is rare, primarily due to a lack of exposed bedrock in the upper reaches of glacier catchments. Previous work from upstream sites $\sim 100\text{km}$ from the coastline suggests limited change in ice surface elevation since the LGM, with some sites documenting thinning during glacial periods (Bockheim et al., 1989; Denton et al., 1989). The thinning history from David Glacier allows for a unique comparison with the modern satellite record and suggests dynamic thinning, can occur $>100\text{km}$ into the interior of the EAIS and can persist over multi-millennial timescales.

The chronologies from David Glacier provide critical insights into glacial processes and rates of change that can be compared against modelled reconstructions from continent scale ice sheet models. Using a data-model comparison software developed by Perry Spector and Greg Balco (dmc.ice-d.org), I extracted a modelled elevation history for David Glacier from four different continental scale ice sheet models and one post glacial rebound model (Argus et al., 2014; Pollard et al., 2016, 2017, 2018). In order to overcome differences in spatial resolution, the scheme extracts an interpolated ice elevation history from the model grid cell containing the field site and it's neighboring model grid cell. Further, most models in the Mt. Kring area have a large variation in true ice sheet elevation suggesting either Mt. Kring has never been ice covered or deglaciated. In order to overcome this artifact, I have applied a bulk shift to the modelled elevation data to ensure the final modelled ice surface elevations lie near the modern ice surface.

The resulting data-model comparison for Hughes Bluff and Mt. Kring reveal a noticeable mismatch in time during the main phase of thinning (Fig 4.8). For Mt. Kring, the main phase of thinning in ice sheet models lags the surface exposure data derived thinning by $\sim 4\text{--}7\text{ kyr}$. For Hughes Bluff, the timing lag is comparable to Mt. Kring with one notable exception being the post glacial rebound model of ICE-6G, likely due to a greater model fit to data constraints along the Scott Coast, particularly multiple relative sea level curves (Baroni and Hall, 2004; Hall, 2009; Argus et al., 2014). The style of thinning is variable between models with noticeable short-lived pauses in thinning, mainly during the Antarctic Cold Reversal ($\sim 15\text{--}13\text{ kya}$) (Pedro et al., 2016). For Mt. Kring, the amplitude of modelled elevation change matches well with the surface exposure age data. As discussed previously, the Hughes Bluff captures a portion of the full thinning event during deglaciation.

Figure 4.8: Surface exposure age data - ice sheet model comparisons for a) Mt. Kring and b) Hughes Bluff. Ice sheet modelled reconstruction data and supporting information from dmc.ice-d.org

During the LGM, ice core records and numerical model outputs suggest widespread ice sheet interior thinning during the LGM while coastal sites show extensive thickening (Verleyen et al., 2011; Mackintosh et al., 2014). The chronology from Mt. Kring, recording ~200 metre thickening followed by monotonic thinning, provides a critical tie point between the high elevation, low accumulation ice domes where ice cores are drilled and the low elevation, high accumulation coastal sites with

more abundant geologic data. Mt. Kring is the first site along the high elevation margin of the EAIS to constrain ice sheet behavior and represents a unique data set for ice sheet reconstructions and continental scale ice sheet modelling experiments.

4.4.2 Terrestrial-Marine Reconstruction

Drawing a link between terrestrial and marine based glacial reconstruction requires a community-based approach to integrate a wide variety of dating techniques, geomorphic interpretations and formation of a common interpretive framework useful for ice sheet modelling efforts (Bentley et al., 2014). The recent effort by Bentley et al. (2014) highlights the utility of a community approach.

During the LGM, nearly all outlet glaciers and grounded ice thickened and expanded to the continental shelf edge surrounding Antarctica (Livingstone et al., 2012; Anderson et al., 2014; Bentley et al., 2014). Since the LGM, the onshore thinning varied spatially but occurred primarily during the Holocene (Small et al., 2019). Along the Transantarctic Mountains, Mid-Holocene outlet glacier thinning and retreat occurred during a relatively stable period of temperature and sea level after the majority of post-glacial sea level rise and rising temperatures occurred (Fig 4.9) (Lambeck et al., 2014; Jones et al., 2015; Spector et al., 2017; Small et al., 2019). Regionally, the synchronous thinning of Mackay, Mawson and David glaciers imply a common control for the observed thinning (Fig. 4.10). For each outlet glacier, our data and modelling suggests subglacial and lateral topography provide the first hindrance to grounding line retreat with a final retreat driven by enhanced sub-ice shelf melt.

Figure 4.9: Palaeo-thinning compilation of Small et al. (2019) updated here to include recent studies of Goehring et al. (2019); Rhee et al. (2019) and this study. See Small et al. (2019) for details on sites 1-23. The majority of records from around Antarctica occur during the early-mid Holocene.

Figure 4.10: Surface exposure age data vs relative elevation for Mackay (Jones et al. (2015), Mawson (Whitmore et al. (2020) and David Glaciers. Broad synchronicity implies a common control for Mid-Holocene thinning.

The David Glacier draws connections between the terrestrial and marine sectors and in order to fully reconstruct the David Glacier, I compare onshore-offshore deglacial chronologies to discuss the synchronicity of marine retreat and terrestrial thinning. Marine geophysical and geological data reveal a large grounding zone wedge (GZW) near Coulman Island (Shipp et al., 1999). Based on marine sedimentary analysis and radiocarbon age dating, the timing of retreat from the Coulman Island GZW initiated at ~ 13 kya, punctuated by few short periods of staggered retreat through the deep Drygalski Trough (Licht et al., 1996; Domack et al., 1999; Anderson et al., 2014). By ~ 10 kya, the grounding line in the Ross Sea migrated south of the David Glacier (Licht and Andrews, 2002; McKay et al., 2008). While these ages are consistent between studies, chapter 6 contains a short discussion on age dating of marine sediments, underlying assumptions, limitations and useful future applications.

Glacial reconstructions using marine geological and geophysical data along the western Ross Sea focus on tracking a grounding line that represents a singular ice sheet (Licht et al., 1996; Domack et al., 1999; Shipp et al., 1999; McKay et al., 2008; Anderson et al., 2014). There is increasing evidence of complex

flow reorganisation and that during deglaciation, palaeo-ice streams and outlet glaciers decouple from the larger ice sheet occupying the central and eastern Ross Sea (Halberstadt et al., 2016; Lee et al., 2017). While it is useful to track the grounding line of the expanded ice sheet in the Ross Sea, as the grounding line retreated past outlet glaciers, local sub-glacial topography is expected to exert a significant control on outlet glacier thinning and retreat (Jones et al., 2015).

Interpreted sub-ice shelf facies from marine sedimentary cores collected in the deepest parts of the inner Drygalski trough yield corrected ages of 4-6 kya (Licht et al., 1996; Frignani et al., 1998; Domack et al., 1999; Licht and Andrews, 2002). This marine evidence may represent the formation of the modern Drygalski Ice Tongue between 4-6 kya, which is consistent with the lowest bedrock age from Hughes Bluff and the raised beach chronology which marks the onset of marine conditions along the Northern Foothills of Terra Nova Bay (Hall et al., 2013). Further, a short pause in glacier thinning may temporally link to short periods of stability recorded by small grounding zone wedges deep in the Drygalski Trough and near Terra Nova Bay (Brancolini et al., 1995; Anderson et al., 2014). Ongoing geomorphic and geologic analysis may provide additional details regarding the timing and nature of this episodic glacial retreat pattern.

4.4.3 Local Oceanographic Impacts

The new geological reconstruction of ice-surface elevation changes at Hughes Bluff shows a rapid lowering of the David Glacier at 6.5 kya and a period of slow thinning from ~6-4 kya. Minimal ice surface lowering since 6 kya likely records the stability of the grounding line within the David Fiord, near the modern configuration, which suggests that the Drygalski Ice Tongue has been stable since ~6 kya. Orombelli et al. (1990) and Baroni and Hall (2004) mapped a series of raised beaches along the Terra Nova Bay coastline that mark beach depositional processes in an open ocean setting (e.g. no grounded ice) initiating at 6 kya. A stable, thinned and retreated David Glacier since 6 kya is consistent with a raised beach chronology along the Terra Nova Bay coastline and suggests a long-term role for Drygalski Ice Tongue in sustaining the Terra Nova Bay polynya as interpreted from modern oceanographic observations (Baroni and Hall, 2004; Stevens et al., 2017).

4.5 Conclusions

Chronologies from the David Glacier reveal a period of glacier thinning along a large swath of the drainage basin during the Mid-Holocene (~6.5 kya). The reconstructed thinning style between two sites separated by ~130km, reveals a

palaeo-dynamic thinning event that endured for two millennia. This chronology is synchronous with local, regional and continental scale geological records of ice sheet behaviour yet is not fully captured in continental scale ice sheet models. In the following chapter, I investigate the potential controls on the Holocene behaviour of the David Glacier basin using a flow line model.

Chapter 5

Flowline modelling of David Glacier

Abstract Understanding the processes and controls on present day Antarctic Ice Sheet mass loss requires knowledge of modern and past ice fluctuations in order to confidently project future sea level rise. The majority of modern outlet glaciers extend seaward as floating ice shelves which act to buttress upstream grounded ice. Ongoing thermal erosion by enhanced ocean heat has the potential to destabilise grounding lines through reduction in ice shelf buttressing. However, observations of grounding line retreat, ice shelf instability and onshore thinning extend back only a few decades. The relatively short period of observations does not cover the full temporal variability of such large ice masses with response times occurring over centennial to millennial timescales. Geological records provide critical insights to past ice sheet change and act to extend the observational record. Numerical modelling of past ice sheet fluctuations, informed by geological records represents a robust approach to investigate the relative role of processes which drove past changes. Here, I model a 1,600 km flowline from Dome C to the Ross Sea continental shelf edge with the aim of disentangling the relative controls on glacier thinning and retreat constrained by geomorphic, geologic and geophysical data. Sensitivity experiments are designed to test the role of enhanced sub-ice shelf melt and lateral buttressing on glacier thinning and grounding line migration. Overall, I show that glacier thinning and grounding line retreat can be initiated by a complex interaction between enhanced sub-ice shelf melt rates and reduction in lateral buttressing which leads to grounding line retreat driven by the Marine Ice Sheet Instability. Once initiated, lateral and sub-glacial geometry controls onshore thinning. For retreat experiments that model grounding line migration to near modern conditions, the resulting onshore modelled thinning closely matches the thinning history constrained by geologic data.

5.1 Introduction, background and justification

Understanding the processes that govern the stability of the Antarctic Ice Sheet is a major focus of ongoing research aimed at assessing the impact of ice mass loss on future global sea level rise (IPCC, 2019). Predictive models, tuned to modern and geological data, provide confidence for projections of future ice sheet behaviour and associated sea level rise yet contain large uncertainties associated with a lack of critical data, representation of key physical processes, and greenhouse gas emission pathways (IPCC, 2013; Golledge et al., 2015; DeConto and Pollard, 2016; Pattyn et al., 2017; Clerc et al., 2019). For example, geological evidence, modern GPS data and numerical modelling studies suggests the irreversible retreat scenario driven by the Marine Ice Sheet Instability (MISI) can be dampened by ice stream lateral buttressing and the solid earth response associated with glacial isostatic adjustment (Gomez et al., 2010; Jamieson et al., 2012; Barletta et al., 2018; Kingslake et al., 2018). Processes associated with ice shelf collapse and grounding line retreat, driven by human-induced, ocean-heat derived sub-ice shelf melt represent the greatest impact on the stability of the marine-based West Antarctic Ice Sheet (Mercer, 1978; Pritchard et al., 2012; Joughin et al., 2012; Dutrieux et al., 2014; Alley et al., 2015; Holland et al., 2019). Further, many processed-based models tuned to modern observational data are run on relatively short timescales and are unlikely to capture longer response times in which ice sheets change (Dutrieux et al., 2014; Dow et al., 2018; Gudmundsson et al., 2019)

The expansion of the Antarctic Ice Sheet (AIS) during the LGM was largely confined by the continental shelf break with the majority of ice volume change occurring in the Ross and Weddell sea sectors (Mackintosh et al., 2011; Golledge et al., 2014; Bentley et al., 2014). Based on geological and geophysical data from marine basins around Antarctica, grounding line retreat initiated first along the Antarctic Peninsula with overall asynchronous retreat behaviour between ~ 18 to 12 kya (Bentley et al., 2014). Broadly, retreat of the AIS from it's maximum position coincides with climate amelioration interpreted from all Antarctic ice cores and deep sea climate record (Lisiecki and Raymo, 2005; Masson-Delmotte et al., 2011; Fudge et al., 2013; Lee et al., 2018).

Focusing on the western Ross Sea and based on marine geological data, grounding line retreat from north of Coulman Island is initiated by ~ 13 ka, with retreat near Terra Nova Bay by ~ 11 ka followed by retreat north of Ross Island by ~ 8.6 kya and a modern configuration by $\sim 2-4$ kya (Anderson et al., 2014; McKay et al., 2016). Onshore, the behaviour of outlet glaciers along the Transantarctic Mountains (TAM) is complex yet largely constrained to the mid-Holocene (Stone et al., 2003; Mackintosh et al., 2007; Todd et al., 2010; Bentley et al., 2014; Jones

et al., 2015; Spector et al., 2017; Small et al., 2019). Outlet glaciers along the Northern Victoria Land sector (Tucker, Aviator, Reeves, Priestley and Campbell Glaciers) initiate thinning at ~ 17 kya with the majority of thinning complete by ~ 7.5 ka, broadly coincident with a linear rise in sea level rise during deglaciation (Baroni and Hall, 2004; Johnson et al., 2008; Smellie et al., 2018; Goehring et al., 2019; Rhee et al., 2019). In contrast, outlet glaciers covering a large swath of the TAM from Southern Victoria Land to Southern TAM, experienced episodic, thinning during the early-mid Holocene, likely due to local topographic effects associated with the Marine Ice Sheet Instability, a process by which ice discharge increases non-linearly due to progressive retreat into landward sloping, deep subglacial basins (Weertman et al., 1974; Mercer, 1978; Schoof, 2007; Jones et al., 2015; Spector et al., 2017). Initiation of thinning at ~ 12 kya along the southern TAM glaciers suggests a complex link with overall grounding line retreat throughout the Ross Sea sector and local topography (Bockheim et al., 1989; Todd et al., 2010; Spector et al., 2017). Overall, variation in geologically constrained thinning behaviour suggests sea-level rise, ice sheet grounding line retreat and overdeepened subglacial topography as general controls for the reconstructed outlet glacier behaviour.

The David Glacier represents a significant element of the Antarctic cryosphere and is the largest outlet glacier in Victoria Land. During the LGM, the David Glacier expanded as an ice stream and coalesced with the grounded ice sheet in the Ross Sea which terminated north of Coulman Island (Licht et al., 1996; Shipp et al., 1999; Livingstone et al., 2012; Anderson et al., 2014). In Chapter 4, I show that the onshore geologically constrained thinning record reveals the David Glacier thinned between 7.5 to 5.5 kya, consistent with local terrestrial records (Stuiver et al., 1981; Orombelli et al., 1990; Baroni and Hall, 2004). The timing of the palaeo-thinning record from David Glacier takes place ~ 3 ka after ice sheet grounding line retreated south of the David Glacier and during a period of sea-level and atmospheric temperature stability at the end of the last glacial termination (Jouzel et al., 2007; Denton et al., 2010; Stenni et al., 2011; Anderson et al., 2014; Lambeck et al., 2014). At Mackay Glacier, modelling indicates that similar rapid thinning and retreat during the Mid-Holocene was driven by MISI, initiated by small changes in sea level rise and/or ocean temperature (Jones et al., 2015).

Along the TAM coast, two lineages of marine geomorphic features exist: one that tracks north-south glacier advance and retreat defined by trough parallel lineations, trough perpendicular grounding zone wedges and one that tracks east-west retreat of local ice towards the coast defined by trough marginal grounding zone wedges (Figure 5.1 and Chapter 3) (Brancolini et al., 1995; Anderson et al., 2014; Halberstadt et al., 2016; Lee et al., 2017). Importantly for the David Glacier,

the inferred intimate relationship with grounded ice in the Ross Sea provides an opportunity to investigate the role of local topographic effects and ice sheet-glacier decoupling on the style of onshore thinning (Jones et al., 2015; Halberstadt et al., 2016). Given the apparent age discrepancy between marine derived grounding line migration and onshore glacier thinning, and the local marine geomorphology, I seek to explore the potential mechanisms for the observed thinning during the mid-Holocene. Using a flow-line model applied to the David Glacier’s geometry, I explore two primary questions: 1) Given the main episode of thinning does not relate to significant increases in atmospheric temperature or global mean sea level, what is the role of ocean heat in driving the observed thinning and retreat? and 2) Given the intimate link between the expanded David Glacier and grounded ice in the Ross Sea, what impact does ice sheet debuitressing have on the timing and style of outlet glacier thinning and retreat?

5.2 Methods

The flowline model used in this study has been described fully elsewhere and has been used in various settings across the Arctic and Antarctic (Viel and Payne, 2005; Nick et al., 2009; Jamieson et al., 2012; Enderlin et al., 2013; Jamieson et al., 2014a; Clason et al., 2016; Whitehouse et al., 2017). Originally designed to track grounding line migration using a moving grid, the model has been used by Jamieson et al. (2012) to test changes in ice stream width and by Nick et al. (2009) and Whitehouse et al. (2017) to incorporate advance treatment of ice shelf dynamics. Here, the model is adapted for David Glacier’s geometry and seeks to explore the relationship between grounding line migration and onshore glacier thinning constrained by surface exposure studies. Background model theory and previous applications are detailed in Chapter 2.

5.2.1 Boundary Conditions

The model domain runs along a $\sim 1,600$ km flowline from Dome C to the continental shelf break (Figure 5.1 and 5.2). Employing ArcGIS tools designed for hydrological modelling, I used modern surface velocity and surface elevation to derive a drainage basin outline and flowline location that follows a zone of streaming ice emanating from the EAIS interior, across the modern grounding line and terminating in Terra Nova Bay (ESRI, 2017; Rignot et al., 2011; Howat et al., 2019). Using bathymetric data compiled by Arndt et al. (2013), the marine sector of the flowline follows the Drygalski Trough axis and runs parallel to mapped mega scale glacial lineations. Onshore, ice stream width follows the drainage

basin outline derived from surface elevation and velocity. Offshore, I employ two methods to determine ice stream width: one by following the 400 mbsl contour consistent with Whitehouse et al. (2017) and another constrained by trough parallel mega scale glacial lineations consistent with Jamieson et al. (2012), herein referred to bathymetrically and geomorphically constrained width, respectively. Generally, the bathymetrically constrained width is wider than the geomorphically constrained width south of Coulman Island by an average of ~ 3 km. North of Coulman Island, sparse multibeam data coverage and fewer MSGs result in similar widths for both methods.

Figure 5.1: Map of model domain with modern surface velocity of (Rignot et al., 2011) and bathymetry of Arndt et al. (2013). Flowline follows red line, ice stream widths in black (solid and dashed).

Modern Boundary Conditions In order to evaluate model performance and appropriate boundary conditions for deglacial experiments, I conducted sensitivity experiments designed to reproduce the modern behaviour of the David Glacier. The model allows the user to input surface and bed geometry, ice stream width, accumulation, velocity and a basal traction parameter. Modern surface elevation, accumulation and ice velocity are well constrained from satellite and *in situ* measurements (Rignot et al., 2011; van Wessem et al., 2018; Howat et al., 2019). In order to account for confluent ice from tributary glaciers I employed an

Figure 5.2: Boundary conditions along flowline distance for deglacial model runs in this study. Inputs include basal topography of Fretwell et al. (2013); Arndt et al. (2013), modelled surface elevation and velocity of Whitehouse et al. (2012) and accumulation of van Wessem et al. (2018). Mapped grounding zone wedges, topographic features, modern day grounding line and terrestrial sites discussed in text are plotted along the upper ice surface and basal topography

ice mass accumulation scheme used in previous modelling studies whereby tributary ice mass is added along the length of the ice stream (Whitehouse et al., 2017). This tributary ‘injection’ is included in the surface mass budget and for the David Glacier model domain, representing ice mass contributions from a secondary ice stream originating from Talos Dome which is observed in surface flow stripe mapping and surface velocity of (Rignot et al., 2011).

Although recent surveys have improved our knowledge of the bed, one key component of the bed has always been well resolved - a prominent bedrock high immediately upstream of the grounding line which currently has a significant stabilising impact on the David Glacier (Morlighem et al., 2019). Further, unpublished airborne gravity inversion modelling of the sub-ice tongue geometry confirms a ~ 500

mbsl bathymetric sill at the mouth of the David Fjord, which is represented in the Bedmap2 compilation (Fretwell et al., 2013). For the reasons outlined above, the final bed geometry input to the model is derived solely from Bedmap2 and the International Bathymetric Chart of the Southern Ocean (Fretwell et al., 2013; Arndt et al., 2013).

Geologic data from former glaciated terrains and satellite observations can be used to approximate basal conditions yet a general lack of data due to inaccessibility of the ice sheet basal environment lead to enduring uncertainties in the basal stress regime (Joughin et al., 2006; Stokes et al., 2015). Based on existing knowledge of sea floor morphology and consistent with previous modelling experiments, the basal traction parameter used in this model attempts a first order approximation of two environs: relatively high basal traction for onshore ice flow over bedrock and relatively low basal traction for offshore ice flow over soft sediments associated with basal till deposits. Modelled basal shear stress values are 20-30 kPa over ice sheet interior and ~ 10 kPa in the marine sector of the flowline which fit with estimates from shear stress measurements on weak subglacial till deposits of Dowdeswell et al. (2004); O Cofaigh et al. (2005) and remote sensing techniques of Vaughan et al. (2003); Joughin et al. (2006).

Modern day sensitivity experiments aim to tune user defined parameters such as internal ice temperature, sub-ice shelf melt rates, basal traction parameter and ice stream width until a best-fit with modern conditions is achieved. Overall, the modern grounding line position, ice shelf thickness and ice shelf length are stable over a 2 kyr modelled period. In contrast, as the David Glacier descends from the ice sheet interior the model was initially unable to reproduce stable conditions. In this zone, the modelled upper surface is steep and undergoes thinning throughout the modelled period. In an effort to stabilise the upper surface upstream of the grounding line, I tuned the basal traction parameter, effectively stiffening the bed, to reduce the modelled instability. Modelled estimates of basal shear stress approach 100 kPa in this setting. Using this set of optimised user-tuned parameters, modern-day modelled output shows good agreement with observations of ice thickness, upper surface elevation and ice velocity.

Deglacial Boundary Conditions The model inputs for deglacial experiments include a reconstructed upper ice surface, bed elevation, ice stream width, accumulation, velocity and basal stress (Figure 5.2). For deglacial experiments, I use the modelled ice thickness and velocity from the geologically constrained continental scale glacial isostatic adjustment model, W12 (Whitehouse et al., 2012) as an initial condition. Using the W12 modelled ice thickness and modern bed elevation allows for an estimate of ice surface elevation without introducing uncertainty involving variable along flow isostatic response and dynamic topography

associated with the long-term evolution of the Antarctic subglacial topography (Stern et al., 2005; Whitehouse et al., 2019; Paxman et al., 2019). The resulting ice surface elevation is consistent with all existing geological constraints (Chapter 4 and serves as an initial ice profile for deglaciation sensitivity experiments.

In order to account for environmental changes during deglaciation, transient changes in accumulation and internal ice temperature are tuned over the modelled period to ensure a stable LGM configuration consistent with geological constraints. Knowledge of past accumulation over glacial-interglacial cycles are restricted to ice core data and internal ice sheet layer mapping near high elevation ice domes (Siebert, 2003; Frezzotti et al., 2005; Buiron et al., 2011; Cavitte et al., 2018). Generally, modern accumulation patterns show relatively low accumulation over the East Antarctic interior and high accumulation near the coastline (Arthern et al., 2006; Lenaerts et al., 2012; van Wessem et al., 2018). Interpretations from ice core records suggest LGM accumulation rates were lower than modern accumulation rates and generally correlates well with temperature proxy records over the Holocene (Siebert, 2003; Veres et al., 2013). I use a scaling relationship between modern accumulation patterns and that accumulation at ~ 15 kya was roughly 75 percent of modern accumulation (Veres et al., 2013). Similar to modern sensitivity experiments, deglacial experiments employ a tributary ‘injection’ scheme and an optimised along flow distribution of the basal traction parameter. Optimised parameter values used to define glacial and interglacial conditions are used as a set of reference forcings that ensure a stable grounding line position and upper ice surface during a 7.5 kyr model spin up period and are designed to represent changing environmental conditions during deglaciation.

In the case of sea level, independent sensitivity experiments were designed to test the impact of gravity and deformation-induced sea-level changes local to the grounding line (Gomez et al., 2010). For these experiments, I forced the model with two modelled relative sea level curves constrained by the local RSL data from Terra Nova Bay (Baroni and Hall, 2004; Briggs and Tarasov, 2013; DeConto and Pollard, 2016).

5.2.2 Deglacial Experiment Approach

Sensitivity experiments designed to understand the controls on thinning and retreat are initialised with a set of user defined parameters derived from modern model runs. Using an optimised set of accumulation and temperature forcings, I explore transient changes in lateral buttressing and sub-ice shelf melt rate to isolate their relative influence on glacier thinning and retreat. Then, I combine the forcings to simulate feedbacks between ocean driven ice shelf thinning and ice

Exp #	Melt Rate	Lat. Buttress Reduction (%)	Resultant Behaviour	
			Geomorphic Width	Bathymetric Width
S1	-1 m/yr	0 / 28	Rapid retreat rate to Sill	Stable CI GZW
S2	-1 m/yr	0 / 31	Rapid retreat rate to Sill	Retreat to modern
M1	-1 / -5 m/yr	0	Moderate retreat rate to sill	Stable south CI GZW
M2	-1 / -9 m/yr	0	Moderate retreat rate to sill	Retreat to modern
M3	-1 / -11 m/yr	0	Moderate retreat rate to modern	Retreat to modern
MS1	-1 / -5 m/yr	0 / 22.5 / 0	Moderate retreat rate to modern	Stable south CI GZW
MS2	-1 / -5 m/yr	0 / 26 / 0	Moderate retreat rate to modern	Retreat to modern
MS3	-1 / -11m/yr	0 / 26	Rapid retreat to modern	Retreat to modern

Table 5.1: Sensitivity experiments in this study with generalised resultant grounding line behaviour for each ice stream width. x / y indicates spin up and forced values, respectively. CI GZW= Coulman Island grounding zone wedge

shelf buttressing recently published by Gudmundsson et al. (2019). All experiments run for 15 kyr with an initial spin up period lasting 7.5 kyr at which point a forcing is applied. For individual forcing scenarios, I linearly decrease lateral buttressing or increase sub-ice shelf melt rate over a 500 year period. In high melt cases, I randomly vary sub-ice shelf melt rate within 0.5 m/yr window to simulate short-lived pulses of warmer or cooler water mass changes.

A set of experiments combining the forcings requires lower threshold values to produce modern surface velocity and geometric relationships between the grounding line location and upper ice surface. In two experiments, the ice shelf buttressing forcing is applied for only 2.5 kyr before linearly increasing to full buttressing, a scenario which attempts to simulate enhanced lateral buttressing by the steep walls of the David Fjord. One experiment explores the impact of debuttressing for the entire deglacial period (7.5 to 0 kyr). In the following sections, I describe the results from sensitivity experiments using a) a geomorphically and b) bathymetrically constrained ice stream width (Table 5.1). To initiate retreat in the bathymetrically constrained ice stream width, I progressively increase sub-ice shelf melt rate or decrease lateral buttressing individually to target appropriate values in which to test model sensitivity to variations in ice stream width. Finally, I compare the modelled grounding line retreat and glacier thinning to two onshore sites which provide high-resolution thinning chronologies derived from surface exposure studies.

5.3 Results

Sensitivity experiments are designed to a) reproduce a geologically consistent glacier thinning and retreat behaviour and b) investigate the relative roles of ice shelf geometry and sub ice shelf melt in driving grounding line retreat and glacier thinning. Modelling the present day David Glacier allows optimised tuning of

user defined parameters for subsequent sensitivity experiments aimed at isolating controls on thinning and retreat. Modelled thinning and retreat behaviour and rates from all experiments are compiled in table 5.3 and the following is a detailed review of each model run.

5.3.1 Geomorphically Constrained Model Runs

Using an ice stream width defined by the orientation and distribution of mega-scale glacial lineations (MSG L), I conducted sensitivity experiments that model the thinning and retreat of the David Glacier to a configuration consistent with modern day observations (Figure 5.4). After a 7.5 kyr model spin up period during which forcings are held constant, ice shelf buttressing is incrementally reduced until retreat is initiated. In this case, retreat occurs at a threshold of 28% although it is not until further ice shelf debuttressing to 31% that the grounding line is able to retreat over the prominent sill at the mouth of the David Fjord (Figure 5.6). In both cases, the reduced buttressing forces extremely rapid grounding line retreat to a prominent sill at the mouth of the David Fjord. Further buttressing reduction results in an unrealistically long (100s of kilometers) and thin ice shelf. In these scenarios, the resulting modelled upper ice surface remains above the Hughes Bluff site when pinned to the sill. At Mt. Kring, modelled rapid thinning is synchronous with Hughes Bluff yet results in an unrealistic final surface elevation 100s of meters below observed modern surface elevation. Modelled maximum thinning rates during the main phase of retreat Mt. Kring and Hughes Bluff are 0.66-0.75 m/yr and 7.12-7.79 m/yr, respectively. Maximum grounding line retreat rates are 62-77 m/yr.

Figure 5.4: a) Sub-ice shelf melt rate and lateral buttressing reduction forcings, b) Time-transgressive elevation profile for Mt. Krøyer and c) Hughes Bluff sites with mid-Holocene surface exposure ages d) Evolution of grounding line position with modern grounding line position, bathymetric sill at mouth of David Fjord and mapped grounding zone features.

Focusing solely on sub-ice shelf melt rate, a similar set of experiments are designed to simulate the impacts of enhanced ocean heat on grounding line retreat (Figure 5.7). After a 7.5 kyr spin up period, a threshold sub-ice shelf melt rate of -14 m/yr achieves rapid grounding line retreat behaviour and is consistent with the modern grounding line position. For melt rates between -5 and -14 m/yr, grounding line retreat is rapid but the final grounding line remains pinned to the prominent sill at the mouth of the David Fjord. Modelled surface reconstructions place the upper surface above the Hughes Bluff site yet agree well with Mt. Kring data constraints. In the high melt case, grounding line retreat stagnates at the sill for approximately 5 kyr. The final retreat phase from the sill to a modern configuration correlates with a final surface consistent with modern observations along the Drygalski Ice Tongue. Modelled maximum thinning rates for Mt. Kring and Hughes Bluff are 0.26-0.44 m/yr and 1.57-2.37 m/yr, respectively. Maximum Grounding line retreat rates are 163-191 m/yr.

Overall when combining forcings, lower threshold values are required to initiate thinning and retreat. For the highest sub-ice shelf melt rate case (-11 m/yr) and 26% reduced lateral buttressing (held for 7.5 kyr), the grounding line behaviour closely matches the high melt only experiment with one notable exception: a second, equally rapid thinning event associated with grounding line retreat to near modern position. Modelled upper ice surface reconstruction agrees well with the Mt. Kring site data yet Hughes Bluff remains ice covered until grounding line retreat from the sill approximately 5 kyr after the main phase of thinning and retreat. For the combined forcing experiments, modelled maximum thinning rates for Mt. Kring and Hughes Bluff are 0.43-0.52 m/yr and 2.45-3.92 m/yr, respectively. Two phases of equally rapid maximum grounding line retreat rate for the three experiments vary from 166-345 m/yr.

5.3.2 Bathymetrically Constrained Model Runs

Following the approach from previous modelling studies of Jamieson et al. (2012, 2016), I constrained an alternate ice stream width by following the 400 mbsl contour and used this as input for sensitivity experiments. Using the same experiment set as with the geomorphically constrained ice stream width, I explore the individual and combined forcings role in driving glacier thinning and retreat (Figure 5.5).

Figure 5.5: a) Sub-ice shelf melt rate and lateral buttressing reduction forcings, b) Time-transgressive elevation profile for Mt. Kring and c) Hughes Bluff sites with mid-Holocene surface exposure ages d) Evolution of grounding line position with modern grounding line position, bathymetric sill at mouth of David Fjord and mapped grounding zone features.

Perturbing only ice shelf buttressing, a 28% reduction in lateral buttressing results in a minor grounding line retreat yet overall stable position south of Coulman Island (near a mapped GZW) (Figure 5.6). A 31% reduction in lateral buttressing forces an extremely rapid grounding line retreat to modern configuration with a short-lived (~ 500 yr) pause at the David Fjord bathymetric sill. Overall, the timing of glacier thinning and retreat lags the forcing by ~ 2 kyr. Final modelled elevations at Hughes Bluff closely match modern observations yet near Mt. Kring, thinning endures throughout the model run resulting in modelled elevations over 500 m below modern observations. In the full retreat scenario, maximum thinning rates for Mt. Kring and Hughes Bluff are 0.96 and 4.22 m/yr, respectively while grounding line retreat rate exceeds 1 km/yr.

When compared to ice shelf debuttressing forcing, experiments forced only by sub-ice shelf melt rate show a distinct timing difference for modelled glacier thinning and retreat (Figure 5.7). For low melt rates (-5 m/yr), grounding line retreat is minimal and remains near Coulman Island. For moderate to high melt rates (-9 to -15 m/yr), grounding line retreat to a modern configuration is rapid with a multi-decadal pause at the David Fjord bathymetric sill. While the behaviour is similar for moderate to high melt forcing, a ~ 1 kyr lag in grounding line retreat occurs in the moderate melt case. Modelled thinning rates for Mt. Kring and Hughes Bluff are 1.15-1.68 m/yr and 4.26-6.22 m/yr, respectively. Grounding line retreat rates are ~ 400 m/yr in both retreat scenarios.

Overall, combining the forcings produces a similar rapid glacier response as with the melt driven case yet requires lower sub-ice shelf melt rates and less lateral buttressing reduction (Figure 5.8). Grounding line retreat and thinning to modern positions occurs with -5 m/yr sub-ice shelf melt rate and a 2.5 kyr, 26% reduction in lateral buttressing. While modelled upper ice surface reconstruction matches well with data from Hughes Bluff, a significant mismatch at Mt. Kring endures. Overall, thinning and retreat rates are very high. Mt. Kring and Hughes Bluff thin at their highest rates of all model runs at 1.64-2.10 m/yr and 7.36-8.6 m/yr, respectively. Grounding line retreat varies from 498-608 m/yr.

5.3.3 Sea Level forced Experiments

A subset of experiments focused on understanding the impact of sea level on grounding line retreat and glacier thinning (Table 5.2). Forced by a range of modelled relative sea level changes through time, I find that the resultant simulations are insensitive to sea level rise. The patterns and timings are similar for unforced simulations (Figures 5.9 and 5.10). Modelled onshore thinning only occurs when combined with an enhanced ice shelf melt rates and a reduction in lateral buttress-

Figure 5.6: Transient flowline profile from all experiments forced only by a reduction in ice stream lateral buttressing.

Figure 5.7: Transient flowline profile from all experiments forced only by an increase in sub-ice shelf melt rate.

Figure 5.8: Transient flowline profile from all experiments forced by a combination of reduction in ice stream lateral buttressing and sub-ice shelf melt rate.

ing (cases RSL4B,5B). As with deglaciation experiments, the modelled pattern on thinning and retreat is sensitive to ice stream width.

Exp #	Temp	Accumulation	Melt	Shelf	Resultant Behaviour			
					RSL80	BathyWidth	RSL40	BathyWidth
RSL1	Temp30	ACC	Melt1	Shelf1	Immediate slow GL retreat to sill	Stable at CI GZW	Immediate slow GL retreat to sill	Stable at CI GZW
RSL2	Temp	ACC	Melt1	Shelf1	Immediate slow GL retreat to sill	Stable at CI GZW	Immediate slow GL retreat to sill	Stable at CI GZW
RSL3	Temp30	ACC1	Melt0	Shelf1	Immediate slow GL retreat to sill	Stable at CI GZW	Immediate slow GL retreat to sill	Stable at CI GZW
RSL4	Temp25	ACC2	Melt13	Shelf3	Immediate slow GL retreat to modern	Rapid retreat to modern	Immediate slow GL retreat to modern	Rapid retreat to modern
RSL5	Temp25	ACC2	Melt13	Shelf6	Immediate slow GL retreat to modern	Rapid retreat to modern	Immediate slow GL retreat to modern	Rapid retreat to modern

Table 5.2: Experiments aimed at understanding model sensitivity to relative sea level. Four sets of simulations are derived from testing two modelled relative sea level curves, each simulated with geomorphically and bathymetrically constrained ice stream widths.

Figure 5.9: Model results from experiments forced by a low relative sea level estimate.

Figure 5.10: Model results from experiments forced by a high relative sea level estimate.

5.4 Discussion

The results of flow line modelling experiments along the expanded David Glacier reveal a threshold-driven sensitivity to both sub-ice shelf melt rate and ice stream lateral buttressing. Further, the modelled thinning and retreat rate is dampened by a narrower ice stream width, constrained by mega-scale glacial lineations, along the marine portion of the flow line. Periods of modelled grounding line retreat are correlated to periods of onshore thinning constrained by surface exposure studies at two locations along the flowline. The results show the glacier response occurs at lower threshold values when the processes act in combination, consistent with previous applications of this model (Jamieson et al., 2014a).

While it is not the intention of this work to reproduce the exact timing of marine retreat, model results provide insights to the dominant processes responsible for glacier thinning and retreat. Although the sensitivity experiments are inconclusive as to which process-geometry combination forced the observed onshore thinning, in the following sections, I explore potential explanations relating to the thinning profile at Hughes Bluff, data-model mismatch at Mt. Kring and a lack of modelled slowdown at all grounding zone wedge locations

5.4.1 Coastal thinning

Using continental scale ice sheet model and available geological data as a guide, the thinning profile at Hughes Bluff captures the final 200 metres of the approximately 600 metre thinning since the LGM (Chapter 4, Figure 4.8 and 3.13). Considering the two ice shelf width geometries used to model the David Glacier in this chapter, only the geomorphically constrained ice stream width geometry produces a long-lived (2.5 kyr) stable grounding line pinned on David Fjord bathymetric sill while maintaining an upper ice surface above Hughes Bluff, consistent with the Hughes Bluff chronology (Chapter 4, Figure 4.4). The subsequent retreat to a near modern configuration reproduces both the magnitude and rate of onshore thinning derived from surface exposure data. During the period of relative grounding line stability, the modelled upper ice surface lies below the upper D’Urville Wall. If the upper surface was below the upper D’Urville Wall, then surface exposure ages of collected erratics should record an older phase of glacier thinning relating to a grounding line retreat from Coulman Island to the David Fjord bathymetric sill (e.g. early-Holocene). See Chapter 6 for more discussion on future work.

5.4.2 Thinning in the ice sheet interior

The modelled thinning profile from Mt. Kring is sensitive to grounding line migrations from the outer and inner continental shelf. While modelled thinning persists during stable grounding line positions (at the bathymetric sill and modern positions), the data-model mismatch in thinning rate is likely due to complex topography and long-distance projection of Mt. Kring to the flowline location. The Mt. Kring site is part of a broad bedrock platform along the northern flank of the major glacial trough dissecting the TAM. The platform is comprised of at least three individual high elevation outcrops protruding ~ 200 metres above the local ice surface, with Mt. Kring nearest to the zone of streaming ice with surface velocities ~ 100 m/yr. Geophysical characterisation of the northern TAM suggests the presence of individual tectonic blocks bounded by faults which likely serve as zones of relatively weak rock strength allowing preferential ice flow (Salvini and Storti, 1999; Dubbini et al., 2010). This tectonic control on long-term ice flow steering is linked to modern inland thinning elsewhere in Antarctica and may act to force complicated dynamic behaviour in the upper David Glacier area (Bingham et al., 2012).

Along the upper reaches of the David Fjord, there are no outcrops in the zone of modern high surface velocity (>100 m/yr). The closest outcrop to fast-flowing ice, Mt. Kring, lies ~ 40 km from the modelled flowline. Modern surface velocities near Mt. Kring are $\sim 15\%$ of the surface velocities at the projected flowline position (Rignot et al., 2011). As an analog, modern satellite derived estimates of glacier thinning are significantly reduced in areas of slow flow with high velocity zones thinning at 0.84 m/yr versus 0.12 m/yr for marginal areas with slower flow velocities, which is consistent with thinning rates in this study (Pritchard et al., 2009). Using modern satellite estimates for glacier thinning as an analog, the Mt. Kring site likely reflects marginal thinning rates rather than along the trunk of the ice stream. Ice loss due to enhanced flow, known as dynamic thinning, is essentially unconstrained over centennial to millennial timescales (Shepherd et al., 2001; Wingham et al., 2009; Pritchard et al., 2009, 2012). Taken together with modern analogs, the Hughes Bluff and Mt. Kring chronologies are interpreted here to record ~ 2 kyr of dynamic thinning along the David Glacier. Overall, the complex topography along the margins of streaming ice flow exerts a significant control on local thinning signals derived from surface exposure studies. Broadly, the Bedmap2 compilation compares well with recent airborne and satellite based estimates of bed elevation along the modelled flowline yet recent improvements in subglacial topography should encourage higher resolution three dimensional modelling for this catchment (Fretwell et al., 2013; Morlighem et al., 2019; Park et al.,

2018).

5.4.3 Controls on grounding line migration

Overall, the style of modelled grounding line retreat is controlled by the Marine Ice Sheet Instability, a feedback process linking enhanced ice discharge, ice sheet thinning and grounding line retreat into progressively deeper subglacial basins (Weertman et al., 1974; Mercer, 1978; Schoof, 2007). The topographic profile of the western Ross Sea is dominated by the deep and landward sloping Drygalski Trough and as a result of increasing discharge during grounding line retreat into deeper waters, the grounding line is modelled to stabilise on a bathymetric sill at the mouth of the David Fjord. Experiments designed to simulate glacier-ice sheet debuddressing show that once independent, the David Glacier grounding line rapidly retreats through the Drygalski Trough to a stable position near a modern configuration. Geologic and geomorphic evidence for glacier-ice sheet coupling is well constrained for ice sheet advance scenarios yet retreat scenarios often attribute grounding line migrations to the broader ice sheet as opposed to individual, larger glaciers such as David Glacier (Licht et al., 1996; Shipp et al., 1999; Domack et al., 1999; McKay et al., 2008; Anderson et al., 2014). A recent ice sheet retreat model of Halberstadt et al. (2016) suggests stagnant ice lingering on higher elevation banks adjacent to large bathymetric troughs. This lingering ice likely interacted with David Glacier ice and impacted lateral buttressing during overall retreat. This scenario is supported the bathymetrically controlled retreat model of Halberstadt et al. (2016). Therefore, thinning and retreat histories of grounded ice resting on bank tops, influence ice stream lateral buttressing. Currently, knowledge of this complex lingering ice history is temporally poorly constrained as the majority of research has focused on trough axes (Anderson et al., 2014; Halberstadt et al., 2016). A scenario reflecting this complexity originates from depositional environment models which suggest a ‘calving bay’-like environment dominated during David Glacier grounding line retreat (Domack et al., 1999, 2003; Leventer et al., 2006). In this scenario, retreat is rapid along the trough axis while lingering ice remains along the lateral margins of the ice stream/glacier. In addition to sedimentary evidence, the abundant large ice berg keelmarks seaward of the Coulman Island GZW provide evidence for the development of a calving bay during deglaciation. Taken together, short-term grounding line stabilisation may be facilitated by complex interaction with other outlet glaciers (Reeves and Priestley glaciers - forming proto-Nansen Ice Sheet/Shelf) of Terra Nova Bay and/or grounded ice lingering on complex topography of the banks surrounding the Drygalski Trough.

In the case of the geomorphically constrained ice stream width, modelled

grounding line migration initially retreats to a location where a large GZW has been documented (Lee, 2019). A lack of modelled grounding line stagnation at other, smaller GZWs are likely indicative of short-lived grounding line stability. Further, in the case of the much smaller ‘GZWs’ observed in the deepest portion of the Drygalski Trough, their morphology may reflect a point source depostional environment versus a line sourced depositional environment associated with more classically defined sheet-like GZWs. Regardless, the small mounds in the trough axis are likely to reflect local stabilising effects during overall grounding line retreat.

5.4.4 Conclusion

In this chapter, I have outlined the flowline modelling approach for the expanded David Glacier. In this approach, I use a well established flowline model to test the relative controls on glacier thinning and grounding line retreat based on on-shore thinning and offshore retreat chronologies. Model results suggest threshold driven control on thinning and retreat by sub-ice shelf melt rate and ice stream lateral buttressing, and that thresholds are reduced when the processes act in combination. Overall, episodes of grounding line retreat correlate well with periods of onshore thinning, constrained by high-resolution surface exposure ages. Data-model mismatches highlight enduring questions related to the relative role of local topographic pinning points, dynamic thinning effects and glacier-ice sheet decoupling. In the following chapter, I seek to synthesise this thesis and offer insights for future research objectives.

Chapter 6

Synthesis

6.1 Response to research questions

This research seeks to answer two questions:

1. *What is the timing and nature of David Glacier thinning since the LGM*
2. *What are the controls on the observed thinning?*

In order to answer these questions, I developed a terrestrial to marine geomorphic framework (Chapter 3) which guided a glacial geologic reconstruction (Chapter 4) and provided context for flowline modelling (Chapter 5) of David Glacier.

Site specific geomorphic characterisation provided the regional to local context required for a surface exposure dating study aimed at tracking the upper ice surface through time. High elevation bedrock surfaces show evidence of a complex exposure history related to burial by cold, non-dynamic ice which minimized glacial erosion and preserved a relict surface. In a regional context, these samples show evidence of glacial overriding and erosion over million year time scales.

Focusing on the LGM to present history, cosmogenic nuclides extracted from bedrock and perched glacial erratics from two sites stretching ~ 150 km along the David Glacier provided evidence for a thinning event initiating at 7.5 ka. At Mt. Kring, a site situated 70 km upstream from the modern grounding line and near a zone of fast flowing ice, a monotonic thinning event is captured over a 2,000 year period. At Hughes Bluff, a low-lying outcrop situated along the Scott Coast at the mouth of the David Glacier, a rapid (up to 2 m/yr) thinning event at approximately 6.5 ka was captured. At Mt. Kring, slower thinning (up to 0.18 m/yr) endured between ~ 7.5 to 5.5 kya. Together, this suggests synchronous glacial thinning along the length of the rapidly flowing portion of the glacier. Subsequent thinning at Hughes Bluff from ~ 6 ka was gradual with

bedrock and erratics displaying the same exposure history which is associated with the emergence of a low bedrock outcrop near the modern surface. The lowest glacial erratic in the study suggests that the modern Drygalski Ice Tongue has remained stable since ~ 4.5 kya.

In order to understand the controls on glacier thinning, I conducted numerical model sensitivity experiments along a 1,600 km glacier flowline which extends from the ice sheet interior to the shelf edge in the western Ross Sea. Following the axis of the Drygalski Trough, the flowline crosses numerous grounding zone wedges of various sizes. Guided by the orientation and distribution of mega-scale glacial lineations as well as bathymetric contours, I determined two separate representations for an idealised palaeo-David ice stream width. The geomorphically and bathymetrically determined widths allowed for sensitivity testing under a set of idealised scenarios which the expanded David Glacier may have experienced during deglaciation of the wider Ross Embayment.

Glacier modelling sought to answer two questions: 1) What is the role of ocean heat in driving onshore glacier thinning and offshore grounding line retreat? and 2) How did David Glacier respond when decoupled from the wider grounded ice in the Ross Sea? In order to answer these questions, sensitivity experiments were designed to explore the response of a stable expanded David Glacier to a range of forcings. This included increasing sub-ice shelf melt rates to simulate an increase in ocean heat to the continental shelf, and decreasing lateral buttressing which may have occurred as grounded ice in the Ross Sea migrated southward of David Glacier. The forcings were combined to explore potential feedbacks associated with the Marine Ice Sheet Instability. Overall, once initiated, the modelled David Glacier retreated rapidly to a stable grounding line position in ~ 500 yrs pausing at a prominent sill at the mouth of David Fiord for up to 5 kyr before subsequent grounding line migration to a modern configuration. This simulated two phase grounding line retreat compares well with geologically observed thinning events at onshore locations along the flowline near two sites (Mt. Kring and Hughes Bluff), both in terms of timing and rates of past glacier thinning. Models forced by moderate sub-ice shelf melt rates (-9 m/yr) fit best with onshore geologic constraints. When forced together, lower melt rates and less lateral buttressing is required to match onshore geologic constraints.

Figure 6.1 and 6.2 synthesise the results of the terrestrial thinning chronology, modelled glacier flowline behaviour and the existing regional marine retreat chronology. Overall, as the grounding line retreats from the prominent bathymetric sill at the mouth of the David Fiord, the David Glacier rapidly thins, decouples from the proto-Nansen Ice Shelf, and allows for open marine conditions to prevail throughout the Terra Nova Bay area.

Figure 6.1: Holocene thinning and retreat model for David Glacier, Terra Nova Bay. Synthesis profile focused on two phase retreat between 11 ka and 5.5 ka. Grounding zone features: well constrained (orange triangles), poorly constrained (green triangles), with relevant studies highlighted (A-G). GL=Grounding line, SIS=Sub-Ice Shelf. Accompanying figure 6.2

Figure 6.2: Map of Holocene thinning and retreat model for David Glacier, Terra Nova Bay. Synthesis map focused on two phase retreat between 11 ka and 5.5 ka, including geographic and geomorphic features mentioned in text. Inset shows provides context for main extent (red box). Inset and main map depict satellite imagery of Bindshadler et al. (2008), bathymetry of Arndt et al. (2013) and surface velocity of Rignot et al. (2011). Accompanying figure 6.1

6.2 Enduring questions and future research

6.2.1 Onset and maximum thickness of David Glacier

Prior to this study, evidence of regional glacial advance and retreat since the LGM was largely confined to the low lying areas of Terra Nova Bay. At the LGM, large outlet glaciers of Terra Nova Bay region (David, Reeves, Priestley and Campbell glaciers) merged and deposited glacial drift across much of the low lying areas of the Northern Foothills and surrounding areas up to 400 meters above sea level (Stuiver et al., 1981; Orombelli et al., 1990). The drift matrix is rich in marine shell fragments that were incorporated when grounded ice overrode the sea floor and deposited the till on land (Orombelli et al., 1990; Hall, 2009). Based on radiocarbon ages from marine shells contained in the drift matrix, the youngest drift (Terra Nova drift) was deposited at 25 to >37 kyr B.P. (Orombelli et al., 1990). Such old radiocarbon ages (approaching 7 half-lives) should be treated with caution due to the detection limit of such small amounts of material (Orombelli et al., 1990). Beach sediments, resting on Terra Nova drift and deposited in parallel ridge sets, occur up to 30 meters above sea level as a result of isostatic uplift following glacial unloading (Priestley, 1923; Stuiver et al., 1981; Baroni and Orombelli, 1991). Over 100 radiocarbon ages of raised beaches provide a record of isostatic uplift of 0.5-2mm/yr over the past 1000 years and up to 8mm/yr between 1-8 kya (Baroni and Orombelli, 1991; Baroni and Hall, 2004). This longer term uplift rate matches well with the vertical component of modern uplift rates derived from local permanent GPS measurements, i.e., 1.91 mm/yr $\pm$ 0.71 (Dubbini et al., 2010). The raised beach chronology has been used by Baroni and Hall (2004) to construct a relative sea level curve which suggests open water conditions (i.e. no grounded ice) since 7.2 kyr and is in broad agreement with the timing of glacial retreat from marine sedimentary records as well as the onshore thinning events identified in this study. (Baroni and Orombelli, 1991; Domack et al., 1999; Licht et al., 1999; Baroni and Hall, 2004).

Two questions remain from the field sites visited in this thesis: 1) How thick did the David Glacier get at the LGM and 2) When did deglaciation initiate? Data presented in this thesis provide the first steps towards answering these questions. In response to the first question, ongoing analysis of bedrock and glacial erratics seek to constrain the coastal David Glacier thickness at the LGM. In collaboration with Carlo Baroni and Maria Cristina Salvatore at University of Pisa and Brent Goehring at Tulane University, we will seek to constrain the upper ice limit by analysing high elevation bedrock and erratic samples from the northern flank of the David Glacier. Existing bedrock samples with ^{10}Be exposure histories greater

than 100 kya from the DUrville Wall and Mt. Neumayer cover nearly 400-800 meters of elevation above the present ice. Some sites show clear rounded summits, presence of erratics and faint glacial striations. Utilising state-of-the-art in situ ^{14}C cosmogenic nuclides, we seek to document the most recent exposure history. Over this elevation range, we expect to see exposure ages within the LGM time frame, allowing refined constraints on minimum ice thickness inferred to have covered these sites. This will also allow further local refinement of the onset of glacier thinning.

An ongoing similar approach is currently underway for samples at Mt. Kring. Here, the youngest ^{10}Be exposure ages show a mid-Holocene thinning history discussed in Chapter 4. Three older exposure age populations suggests that these samples have a complex exposure history (e.g. inheritance). At the time of writing this thesis, glacial erratics in the middle-aged population (51-25 kya) are being analysed for in situ ^{14}C exposure history. The highest elevation samples from this population are likely to be the last covered during glacial periods and the first exposed during the last deglaciation. In this scenario, decay during burial under LGM ice would likely have led to the decay of the previously accumulated ^{14}C such that these samples will only carry a memory of the most recent exposure period (since 7.5 kya) and that a two nuclide burial approach will provide additional constraints on the onset of glaciation in this high elevation, ice sheet interior setting. Further, recent extraction of ^{10}Be from pyroxene in mafic rocks may provide additional constraints on the maximum ice thickness at Mt. Kring (Collins, 2015; Eaves et al., 2018). In this scenario, ^{10}Be extracted from dolerite bedrock which was collected by Carlo Baroni will afford the opportunity to provide additional details on the apparent complex exposure and burial history at the peak of Mt. Kring.

Together, the additional sample analysis in combination with emerging methods will provide new and refined answers to questions related to the onset and maximum ice thickness at the LGM for this sector of the Antarctic Ice Sheet. This represents a significant opportunity for further understanding of David Glacier's history that this current study does not address fully.

6.2.2 Glacial reconstruction in the western Ross Sea

The history of marine-based grounded ice retreat in the western Ross Sea starting at 11 ka, is constrained by the radiocarbon age of ice-proximal glacial-marine sedimentary facies from marine sediment cores, seismic stratigraphy and interpretation of submarine geomorphic features (MSGs and GZWs) (Licht et al., 1996, 1999; Domack et al., 1999; Shipp et al., 1999). Interpretation and age control

from marine sediment cores help to constrain grounded ice history around the Antarctic continental margin, however, sparse data, uncertain Antarctic marine reservoir effects, and corresponding large age uncertainties are limiting factors in obtaining a coherent chronology of ice extent and grounding line retreat (Andrews et al., 1999). There is a long history of using the acid-insoluble organic (AIO) fraction of diatomaceous mud for age-depth calibration, but the core tops are typically much older than the 1200 ^{14}C yr ocean reservoir correction determined for marine carbonates (Andrews et al., 1999). The reworking and deposition of ‘dead carbon increases the measured ^{14}C age representing a significant limitation for using AIO ages from marine sediment cores (Andrews et al., 1999; Licht et al., 1999; McKay et al., 2008). Dating marine sediments using carbonate-bearing shell fragments found in proximal glaciomarine facies is preferred but such material is rare (Domack et al., 1999; McKay et al., 2008, 2016; Hillenbrand et al., 2017).

Ongoing marine geological and geophysical mapping seeks to constrain former ice extent, source area, and glacial retreat style and timing. I have organised three Scientific Committee on Antarctic Research workshops aimed at developing a terrestrial to marine, community-based, glacial reconstruction in the wider Terra Nova Bay region. This collaborative work has already enhanced our collective understanding of data gaps, identified newly collected data sets, and helped motivate the application of novel approaches to reduce errors associated with dating marine sediments. Use of ramped pyrolysis of radiocarbon, emergence of novel applications of biomarkers in marine sediments, incorporation of a growing set of marker beds from cryptotephra analysis, as well as improved coverage of multibeam and high frequency seismic data will allow a next generation of glacial reconstruction in the marine sector.

This SCAR workshop group has allowed enhanced collaboration with onshore glacial geology groups led by Carlo Baroni (University of Pisa) and Yeong Bae Seong (Korea University). Through this collaboration and combining field experience and unpublished data, we have identified gaps in the onshore record in which to focus future efforts. Additional constraints on tributary glaciers (Reeves, Priestley and Campbell Glaciers) will produce a more robust glacial reconstruction particularly given the incredible array of submarine geomorphic indicators which suggest an intimate link with the expanded David Glacier.

6.2.3 Complex flow of David Glacier

The modelling results presented in this study provide initial, simplified process-based understanding of the expanded David Glacier. This study is not without limitations which may be overcome through additional efforts related to improved

mapping of the bed elevation and computational efficiencies which may allow for incorporation of other processes not considered in this study. Recently enhanced coverage of bed elevations of Morlighem et al. (2019), as well as ongoing subglacial and submarine mapping will further refine bed topography allowing for improved 3D flow modelling. The model must be capable of robust grounding line tracking as well as fine enough resolution to simulate smaller valley glaciers of the Terra Nova Bay area. Additionally, the modelling approach in this study does not incorporate processes associated with subglacial water flow, isostatic rebound or represent a polythermal glacial system which are factors that may lead to more robust modelled reconstructions.

Given the size of David Glacier and the multiple tributary glaciers both near the modern grounding line and within the LGM expanded footprint into the marine setting suggests the importance of improved treatment of confluent ice. In particular, the modern David Glacier flow is dominated by ice emanating from Dome C with a significant contribution from Talos Dome. Both of these tributaries have ice cores at the margins of the drainage basin and represent much different accumulation regimes (van Wessem et al., 2018). Talos Dome is a more coastal site which experiences greater accumulation than the high polar desert environment of Dome C. Geologic constraints on past glacial behaviour along these tributaries may reflect this difference in accumulation and could be explored further through analysis of two additional sites introduced in this study. Western Trio Nunataks and Evans Heights are two sites which lie immediately adjacent to the two main flow units feeding into the David Fiord. Their thinning histories, potentially constrained by high-resolution, low-inheritance surface exposure ages consistent with Hughes Bluff and Mt. Kring may help unravel any complexities associated with differential accumulation patterns from their respective ice domes. In fact, in Chapter 4, I introduce the lowest recorded ^3He exposure age in Antarctica and this provides confidence that a chronology from Evans Heights will yield a Holocene aged thinning chronology.

6.3 Conclusion and contributions

This study has made several clear contributions to the broader understanding of the Antarctic Ice Sheet. Through careful collection of glacial deposits from numerous sites along the David Glacier, I have closed a spatial and temporal gap in the rapidly expanding onshore glacial geologic knowledge bank. The data sets generated in this thesis constrain one of the largest outlet glaciers in the world which carries significance for its specific sector as well as offering clues about processes currently underway in rapidly changing sectors in Antarctica and

Greenland today. In particular, if the data and modelling presented in this thesis are representative of outlet glacier behaviour more generally, the ice sheet thinning observed by satellites over the last ~ 40 years may be the first signs of a millennial-scale response.

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