Vulnerability of the Ross Ice Shelf: Seismic Site Characterisation and Drilling Recommendation

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a thesis submitted for the degree of

Master of Science

at the University of Otago, Dunedin, New Zealand

31st March 2016
Abstract

The Ross Ice Shelf (RIS) is the largest ice shelf in the world. It buttresses ice flow from both the East Antarctic Ice Sheet (EAIS) and the West Antarctic Ice Sheet (WAIS). Today the RIS does not appear to be retreating or advancing. Understanding what controls the ice shelf's stability, and how it may respond to future warming and oceanic change is vital, as its collapse would accelerate global sea level rise. Understanding the rates at which physical processes occurred in the past during ice shelf and ice sheet retreat can improve our models for future climate change.

This study aimed to answer two main research questions: first, to characterise seafloor bathymetry and substrates in the vicinity of a future hot water drill site and make informed decisions concerning seafloor coring/sampling locations, and second, to characterise the roughness of the ice shelf’s basal surface.

Field work was conducted during the 2015/2016 Antarctic field season as opposed to the 2014/2015 season, after a one year delay due to logistical constraints. As a result, this study focussed on reprocessing previous data, survey design, and modelling, conducted prior to data collection in Antarctica. The preliminary work involved a comparative study between conventional spiked geophone data and snow streamer on data previously collected on the McMurdo Ice Shelf (MIS), and a detailed survey design for the November 2015 survey. Synthetic shot records were generated to test the effect of ice thickness variations. The snow streamer and weight drop seismic source data acquisition system were an effective method of data collection on the RIS. This combined system allowed for rapid data collection, and facilitated the collection of 45.8 km of multichannel seismic reflection data.

The seismic data are interpreted to reveal two seismic facies, separated by an erosion surface, of at least 180 m thickness. The upper seismic facies is characterised by two cycles of high-amplitude, mostly continuous, horizontal strata, and the lower facies is characterised by irregular, discontinuous, dipping strata. The two seismic facies and
erosion surface are interpreted to reflect the change in glaciation regime that occurred in the late Pliocene (approximately 3 Ma), where the lower sedimentary packages consists of sediments deposited under a warmer, wet-base regime and overlying sediments that were deposited by colder, dry-base glaciers. It is unlikely that deeper bedrock structures were imaged in this study. From the seismic data alone, it is recommended that any future hot water drill site locations are positioned close to the South Pole Overland Traverse (SPOT) road and the 2015 season base camp. The basal ice interface was not imaged distinctly in this study, likely due to the interference of surface waves and the presence of marine ice. It lies within as a seismically opaque zone in the upper 200 ms, after which the signal changes character to low- to moderate-amplitudes in the water column. It is hypothesised that this is due to either the presence of marine ice, surface waves obscuring the reflection, or a combination of the two. The RIS data also display a relatively strong intra-ice multiple (modelled in the synthetic shot records), and contain strong surface waves, which were a significant aspect of the shot records.

Due to the nature of collecting data close to the end of this study, several processing and analysis options still need to be investigated for these data including, but not limited to, better analysis of the surface waves and of the intra-ice multiple characterise ice properties, and calculations of reflection and transmission co-efficient values derived from the intra-ice multiple and seafloor.
Acknowledgements

Firstly, I would like to thank my supervisors Dr. Christian Ohneiser and Dr. Andrew Gorman for their help and advice throughout the duration of my research. They have both been incredibly supportive, their enthusiasm has been fantastic, and the field work just would not have been the same without them! Next I would like to thank Dr. Christina Hulbe who has been very helpful, and always willing to give advice when I needed it.

Next, I would like to thank my family, in particular my parents. They have been incredibly supportive of me throughout my many years of university, especially so in this last year, and I could not have done this without them. The rest of my family and friends have been fantastic during this year, and a special thanks goes out to them.

My research would not have been made possible without the support of multiple agencies, and I am very grateful to the New Zealand Antarctic Research Institute, Antarctica New Zealand, the Alfred Wegener Institute, the Earthquake Commission, the Polar Environments Research Theme at the University of Otago, the Incorporated Research Institutions for Seismology and Portable Array Seismic Studies of the Continental Lithosphere, and GNS Science for their GLOBE Claritas data processing software.

Lastly, I would like to thank the staff and students within the Geology Department, for their help and advice during the year, and the staff from Antarctica New Zealand for their support while at Scott Base and in the field.
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<td>Antarctic bottom water</td>
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<td>AGC</td>
<td>Automatic Gain Control</td>
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<td>AI</td>
<td>Acoustic impedance</td>
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<td>ApRES</td>
<td>Autonomous phase sensitive radio echo sounding</td>
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<td>ANDRILL</td>
<td>Antarctic Geological Drilling Programme</td>
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<td>CDP</td>
<td>Common depth point</td>
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<td>CDW</td>
<td>Circumpolar deep water</td>
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<td>f-k</td>
<td>Frequency-wavenumber</td>
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<td>GLONASS</td>
<td>GLObal NAvigation Satellite System</td>
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<tr>
<td>GNSS</td>
<td>Global Navigation Satellite System</td>
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<tr>
<td>GPS</td>
<td>Global Positioning System</td>
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<td>GSE</td>
<td>Glacial surface of erosion</td>
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<td>HSSW</td>
<td>High-salinity shelf water</td>
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<td>Hot water drill site</td>
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<td>IG</td>
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<td>IPCC</td>
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<td>ISW</td>
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<td>JCS</td>
<td>Job control system</td>
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<td>LGM</td>
<td>Last Glacial Maximum</td>
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<td>MIS</td>
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<td>NMO</td>
<td>Normal moveout</td>
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<td>NZARI</td>
<td>New Zealand Antarctic Research Institute</td>
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<td>RCP</td>
<td>Representative concentration pathways</td>
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<td>RIGGS</td>
<td>Ross Ice Geophysical and Glaciological Survey</td>
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<td>RIS</td>
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<td>SMS</td>
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<td>South Pole Overland Traverse</td>
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<td>TAM</td>
<td>Transantarctic Mountain</td>
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<tr>
<td>t-x</td>
<td>Time-distance</td>
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<td>West Antarctic Rift System</td>
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<td>Weight drop seismic source</td>
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Part I - Introduction and Background
1. Introduction and Background

The Ross Ice Shelf (RIS) is the largest ice shelf in the world. It buttresses ice flow from both the East Antarctic Ice Sheet (EAIS) and the West Antarctic Ice Sheet (WAIS). Today the RIS is neither retreating nor advancing. Understanding the controls on the ice shelf’s stability, and how it may respond to future warming (or cooling) and oceanic change is vital. The collapse of the RIS would accelerate global sea level rise, so understanding the rates at which physical processes occurred in the past during ice shelf and ice sheet retreat can improve our models for future climate change. The RIS programme of the New Zealand Antarctic Research Institute (NZARI) aims to address change in the ice shelf by bringing together scientists from multiple disciplines to collect and analyse new field data over several years.

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1.1 Ice Shelf Structure

Ice shelves can form at the margin of a grounded glacier or ice sheet, such as the EAIS and WAIS, where ice enters the sea, becomes buoyant, and floats as an ice shelf (Drewry, 1986). Ice shelves are complex structures, but can be broken down to three main elements; the upper most firn layer (in accumulation zones, not present in ablation zones), the glacial ice, and the basal interface with the ocean. Ice shelf layers can have a dramatic impact on seismic data quality because of internal reflections, refractions and attenuation of energy.

1.1.1 Firn Layer

Polar firn is a particular type of snow layer, which is consolidated on top of glacial ice, and has gone through at least one melt season (i.e. during summer months, where individual crystals are near melting point and bond together) (Beaudoin et al. 1992). Consolidation occurs through multiple processes, such as grain reorganisation, pressure melting, and recrystallisation (Wilkinson, 1988; Beaudoin et al. 1992). From the surface downwards, the density of firn increases, until the firn changes into glacial ice where its permeability to liquids becomes negligible, and the pores are closed (Herron & Langway, 1980; Beaudoin et al. 1992). The Antarctic ice sheet and RIS are almost entirely covered in a firn layer, which typically makes up the upper 50 - 100 m (Ligtenberg et al. 2011). Due to the wide range of surface climate conditions in Antarctica, firn layer depth and density can be highly variable, where areas absent of firn (known as ablation areas), can result in the development of blue ice (Figure 1.1; Li & Zwally, 2004; Winther et al. 2001). The density of this firn layer is typically close to 0.35 g/cm³ near the top and densifies under pressure (Thiel & Ostenso, 1961). The firn layer has a strong influence on seismic data, as it can alter the shape of the seismic signal, the frequency content, and introduce strong multiples into the data (Beaudoin et al. 1992). As physical and chemical properties of snow can vary, so can firn vary to adjacent layers (Alley, 1988; Albert et al. 2000).
Figure 1.1 The extent of modelled surface density of firn at 550 kg/m$^3$ (a) and 830 kg/m$^3$ (b), a value close to the density of glacial ice, 900 kg/m$^3$. These thicknesses give an indication of firn thickness on the Antarctic continent. The Ross Ice Shelf (RIS) has a modelled firn thickness of approximately 50 to 60 m at 830 kg/m$^3$ (van den Broeke, 2008).

1.1.2 Glacial Ice

Glacial ice makes up the largest component of the RIS and can be bounded by the firn layer above, and the basal boundary below. Ice is considered to be a mineral and has a polycrystalline structure, where interlocking crystals are evident in naturally occurring glacial ice (Drewry, 1986). Ice crystals grow through very low pressure metamorphism (both sintering and compaction) of firn, where grains display a random orientation in the ice. However, due to stress on the ice and deformation from glacial flow, ice can exhibit a preferential crystal orientation (Drewry, 1986). This preferred crystal orientation, as well as seasonal variation of snow production and accumulation, can produce layering within the glacial ice (Legrand & Mayewski, 1997). Glacial ice may contain air that has been trapped from the sealing-off process at the transition from firn to ice (Drewry, 1986).
Ice layering is largely due to deformation and stress within the ice (Catania et al. 2005; Catania et al. 2010; Hulbe & Fahnestock, 2007; Legrand & Mayewski, 1997). Ground-based radar systems can be used to detect larger scale internal layering, which is largely due to flow deformation and stresses within the ice (Figure 1.2). These deformed, deeper layers are consistent with conditions that would be expected from within active ice streams, and can show century-scale flow variability (Catania et al. 2005; Catania et al. 2010; Hulbe & Fahnestock, 2007). Layers which form from seasonal variation, or over longer periods, are at a much smaller scale (centimetres to metres) than deformational layering, and largely form because of changing atmospheric conditions, which in turn are a product of both climatic variability and human activities (Legrand & Mayewski, 1997). It is unlikely that these layers will be observed in seismic data as they are expressed mostly as geochemical variations rather than physical variations that would lead to changes in acoustic impedance.

![Radar profile across the grounding line at Siple Dome](image)

**Figure 1.2** Radar profile across the grounding line at Siple Dome (white arrow, white box is the margin of error in the grounding line position), with ice flow from left to right. The profile shows internal ice layering caused by flow deformation (Catania et al. 2010).

### 1.1.3 Basal Interface

Antarctica’s ice balance consists of snow accumulation within its interior, and mass loss through the ice shelves into the ocean. Ice shelves play an important factor in stabilising the ice sheet, due to buttressing of the ice sheet. Removal of this buttress
from the system, due to ice shelf loss or thinning, removes the integral structure that slows the ice from flowing into the ocean. This is likely to be triggered by oceanic and atmospheric warming (DePoorter et al. 2013). As approximately 60 to 80% of Antarctic ice passes through the ice shelves, the basal boundary is an important interface between the ocean and the Antarctic ice sheet. At the base of the ice shelf, the ice is exposed to warmer water that has transported heat from the Southern Ocean (Pritchard et al. 2012), where it may be lost through basal melting and calving (Jacobs et al. 1992; Smedsrud & Jenkins, 2004).

The amount of mass lost from basal ice varies greatly from shelf to shelf, ranging from 10% to 90% (DePoorter et al. 2013). The larger ice shelves report the lowest melt rates, contributing to approximately 15% of the yearly ice shelf meltwater, despite making up 61% of all Antarctic ice shelves. This is likely due to the lesser effect oceanic heat has on larger ice shelves, and they are more likely to accrete ice at the base (Rignot et al. 2013). The base of the RIS is melting at an approximate rate of 1 m/year (DePoorter et al. 2013; Figure 1.3). Basal crevasses are also common in ice shelves, and were possibly observed on the RIS (Hulbe et al. 2013). Crevasses increase the surface area available for melting and facilitate heat exchange between the ice shelf and the ocean (Jordan et al. 2014).
Figure 1.3 Melting rates (m/yr) of the ice shelves around Antarctica. Note the relatively low melt amount of the Ross Ice Shelf (RIS) compared to the smaller ice shelves (DePoorter et al. 2013).

The thinning of an ice shelf due to basal melt implies the presence of heat supply from the ocean to the underside of the ice shelf and into the sub-ice cavity (Pritchard et al. 2012). There are three main mechanisms through which warm water circulates underneath ice shelves, enabling basal melting (Figure 1.4; DePoorter et al. 2013). The first mechanism produces high-salinity shelf water (HSSW) that results from brine rejection as sea ice forms (Nicholls & Østerhus, 2004). HSSW intrudes the sub-ice cavity beneath the ice shelf, where it reaches and melts the grounding line. At this stage HSSW mixes with the fresher, melted water from the grounding line, and forms ice-shelf water (ISW) (DePoorter et al. 2013; Holland et al. 2008). As a result, buoyant ISW melt water is colder than the in situ freezing point of the ice shelf base, and the supercooled water plume rises along the base of the ice shelf (Smetsrud & Jenkins, 2004). As the supercooled plume rises, it precipitates marine ice at the ice shelf base, and partially refreezes the ice shelf. As a result, most of the melting along the basal ice shelf occurs at the grounding line and calving front (DePoorter et al. 2013; Rignot et al.)
Circumpolar deep water (CDW) is the second main melting mechanism, which intrudes into the sub-ice cavity (DePoorter et al. 2013). CDW travels along sea floor bathymetric troughs underneath the ice shelves, and can reach temperatures of 4°C. This plays an important factor in basal ice shelf melt (Pritchard et al. 2012). Lastly, tidal and wind mixing of warm water near the ice shelf edge is the third main mechanism that causes basal melt (DePoorter et al. 2013). In addition to these three main melting mechanisms, the interaction between oceanic and ISW produces Antarctic bottom water (ABW), which is a key aspect of the global water mass circulation (Holland et al. 2008). For these reasons, it is not surprising that most literature agrees that as ocean temperatures increase, net basal melting will increase too (e.g. Holland et al. 2008; Pritchard et al. 2012; Rignot et al. 2013; Bintanja et al. 2013).

![Figure 1.4](image)

*Figure 1.4* Schematic of the basic structure of an ice shelf and the associated water masses, showing interactions beneath the ice shelf of circumpolar deep water (CDW), Antarctic bottom water (ABW), high salinity shelf water (HSSW), and ice-shelf water (ISW). Modified from Aguilera (2015).

Frazil ice can accrete to the basal ice shelf in the form of marine ice, (e.g. in sutures, troughs, cavities, and rifts, or as a massive layer below the ice), deposited by the buoyant ISW plume (Koch et al. 2015; DePoorter et al. 2013). Pressure increases with increasing depth underneath the ice shelf, and the freezing temperature decreases. Any water below the ice shelf that is at the same temperature as the surface melting temperature, has the ability to melt the basal ice shelf. The resultant meltwater is fresher, cooler, and lighter than the surrounding saline water (Jordan et al. 2014). As this plume of water rises, the increasing *in situ* freezing temperature can supercool the
meltwater mass (Holland et al. 2008; Jordan et al. 2014; Smedsrud, & Jenkins, 2004). This can produce a suspension of disk-shaped, frazil ice crystals as the decreasing pressure in the supercooled plume increases the freezing temperature, and these can subsequently accrete onto the ice shelf base where water flow is slow enough (Figure 1.5; Bombosch & Jenkins, 1995; Holland et al. 2008; Smedsrud & Jenkins, 2004; Jordan et al. 2014). This suspension of frazil ice is initially deposited as slush at the ice shelf base, and can subsequently consolidate (Smedsrud & Jenkins, 2004). However, it is currently still unclear as to whether the marine ice beneath the ice shelves is formed just from frazil crystals which have been compacted, or whether the pore fluid between the crystals freezes due to the conduction of heat through the ice shelf (Koch et al. 2015).

![Figure 1.5 Diagram of platelet marine ice development. Time progresses from left to right, where initial sea ice formation is given by \( t_{form} \). At \( t_{pt} \) there is the transition between columnar to platelet ice. Towards the end of winter, \( t_{end} \), the vertical column shows the transition between snow, \( H_{snow} \), to granular or columnar ice with incorporated platelet ice, \( H_{pt} \), to the sea ice base, \( H_{ice} \), and the base of the sub-ice platelet layer, \( H_{pl} \). The change from columnar ice to incorporated platelet ice takes place at time \( t_{pt} \) (modified from Langhorne et al. 2015).](image)
1.2 Seismic Data Acquisition on Ice Shelves

The ice shelf environment creates unique challenges during seismic data acquisition which are not present in marine or land seismic surveys. Flexural ice waves, infragravity waves, the firn layer, velocity inversion at the ice-water interface, and marine ice present on the ice shelf base all provide unique problems that need to be taken into consideration when collecting and processing seismic data. The following section briefly discusses each problem and how this might be mitigated, either during data collection or processing.

1.2.1 Flexural Ice Waves

One of the challenges when collecting seismic data on floating ice is the generation of an ice flexural wave, which is unique to the ice environment (Henley, 2005). The flexural wave is a strong, coherent signal that is produced as seismic energy generated from a sound source is trapped within the ice, as opposed to travelling through the water column into sediments. This is predominantly due to the large velocity contrasts between air and ice, or ice and water (Barr et al. 1993; Figure 1.6). Flexural waves have very distinct characteristics, where high frequencies travel at high velocities, close to the P-wave velocity of ice, and low frequencies travel at low velocities, close to the P-wave velocity of air. This characteristic, combined with it being a strong, coherent noise, makes it very difficult to remove or dampen in seismic data which can obscure weaker reflections that arrive at a similar time (Barr et al. 1993; Henley, 2005).

**Figure 1.6** There are two main types of flexural waves, simple flexural (a) and flexural-dilatational (b), where the blue lines are the normal state of the ice, and the black lines the ice in the state of flexing (adapted from Liu & Bhattacharya, 2009).
Changing seismic acquisition parameters, for example by changing station spacing, or modelling the noise in the radial trace domain and subtracting in the space-time domain, can partially attenuate the flexural wave; however, such methods are not usually effective, and the flexural ice wave can remain a problem in ice data (Henley, 2005).

1.2.2 Infragravity Waves

Ice shelves are constantly and significantly subjected to multiple components of ocean waves (Bromirski & Stephen, 2012). Oceanic infragravity waves (IG) are high-amplitude waves that can be generated along the coastlines of continents, and travel trans-oceanically to Antarctica, such as those generated on the North American Pacific coast. At the transferral of this high-amplitude swell energy (several metres) to the RIS, mechanical vibrations and flexural stresses can be produced upon impact (Bromirski et al. 2010; Bromirski & Stephen, 2012; Figure 1.7). This response depends on both the wave period and wave amplitude, and is larger during the austral summer where sea ice does not dampen the swell (Bromirski & Stephen, 2012). Changes in basal ice shelf properties can alter the signal propagation of IG, and the relative difference of the IG impact on ice shelves can be estimated when compared to land-based seismic stations (Bromirski & Stephen, 2012). The IG waves can have the ability to affect the stability of ice shelves through various means, such as fracturing the ice, opening crevasses, and collapse initiation (Bromirski et al. 2010). However, currently there is no model that can describe the propagation of IG comprehensively, and IG can affect data quality and introduce noise into seismic data (Wang & Shen, 2010).
Figure 1.7 Cross-section of a floating ice shelf, shown with schematic representations of possible infragravity wave excitation locations (green arrows in the ice shelf). Modified from Bromirski et al. (2015).

1.2.3 Seismic Wave Attenuation in Firn

Seismic wave attenuation is defined as the “reduction in amplitude or energy caused by the physical properties of the transmitting media or system” (Sheriff, 2002: p. 18). Firn is composed of both snow and air, and it has the ability to easily attenuate seismic energy, particularly near the top of the section where there is a larger component of air present. This affects both the P-wave velocity and density.

P-wave velocity is highly variable within the firn layer, and is an important aspect to consider when processing seismic data (Figure 1.8). The upper 20 m of the firn layer has a very high velocity gradient that ranges from approximately 500 m/s at the surface, 2000 m/s at 10 m, and 2750 m/s at 20 m (Thiel & Ostenso, 1961; Beaudoin et al. 1992). Between 20 to 70 m, the velocity gradient decreases until it reaches the peak P-wave velocity of 3800 m/s. There is no consensus within the literature on the true value of the P-wave velocity in glacial ice, of which the ice condition variability is largely responsible. Currently, the most accepted value ranges between 3400 to 3800 m/s (Mussett & Khan, 2000; Thiel & Ostenso, 1961, respectively). Maximum
velocities are measured between 80 to 100 m depth, and are more consistent with those recorded in glacial ice (Thiel & Ostenso, 1961). The S-wave velocity has been modelled by King and Jarvis (2008) and is one of few S-wave velocity profiles known through firn.

Figure 1.8 Changing P-wave velocity with depth through firn and glacial ice (and one S-wave velocity profile (green)). Black: Thiel & Ostenso, 1961 (ice shelf); Red: Robinson, 1968 (ice shelf and grounded ice); Blue: Beaudoin et al. 1992 (ice shelf); Green: King & Jarvis, 2008 (grounded ice). Adapted from Thiel and Ostenso (1961). Multiple profiles for a single reference are from different coring sites.

Firn density is also important to consider when processing seismic data, as it can affect the attenuation of seismic energy. Firn is typically characterised by a density range between that of surface snow (approximately 350 kg/m$^3$) and that of glacial ice (900 kg/m$^3$) (Thiel & Ostenso, 1961; Ligtenberg et al. 2011). The density of firn can be
affected by several factors, such as the degree to which melting has occurred, snow temperature, burial rate, and wind speed near the surface (Li & Zwally, 2004; van den Broeke, 2008). Densification with depth occurs due packing and recrystallisation of the surface snow as the firn is buried further (King & Jarvis, 2008). At a depth of 15 m, the firn will densify at a constant temperature, and is not influenced by surface temperature (van den Broeke, 2008). This densification can take centuries or millennia in areas of low accumulation and low temperatures (Ligtenberg et al. 2011; Figure 1.9).

**Figure 1.9** Changing firn and ice density, on the x-axis, with depth, on the y-axis. Black: Thiel and Ostenso, 1961 (ice shelf); Blue: van den Broeke, 2008 (ice shelf and grounded ice); Green: King & Jarvis, 2008 (grounded ice); Red: Ligtenberg et al. 2011 (ice shelf and grounded ice). Adapted from Thiel and Ostenso (1961). Multiple profiles for a single reference are from different coring sites.

### 1.2.4 Seismic Wave Attenuation in Glacial Ice

Temperature has been referred to as the most important control of seismic wave attenuation in glacial ice, especially as the temperature nears the ice melting point (Peters et al. 2012). Friction produced within the ice from seismic energy travelling through the medium is converted to heat along any present interfaces. As a result of
melting, semi-liquid films along grain boundaries often dominate seismic attenuation in polycrystalline ice formations such as glacial ice. The least amount of attenuation within these types of ice occurs where the propagation of seismic waves is parallel to the main axis of anisotropy (Peters et al. 2012).

1.2.5 Basal Marine Ice
Seismically, marine ice can be an important component of the system to consider as it can produce a ‘slushy’ zone at the base of the shelf in accumulation areas (Figure 1.5). This can obscure the ice-water interface, as there is no clear property change, but rather a gradual change (Frankenstein et al. 2001). The lack of distinct change at an interface results in little or no seismic reflection at the interface, often making it transparent in seismic data (Mussett & Khan, 2000). For these reasons, the presence of marine ice will hinder the imaging of the ice shelf base.

1.2.6 Velocity Inversion
A problem with seismic data collected on ice shelves is the deviation from a ‘normal’ velocity model where P-wave velocity is low at the surface and increases with depth. In an ice shelf environment velocities have a large range in the firn (1000 to 3400 m/s; Figure 1.10), high in the glacial ice (approximately 3400 to 3800 m/s), then lower in the water column (1500 m/s), to again higher in sediments (approximately 1700 m/s). When seismic waves refract from a high velocity layer \(v_1\) into a low velocity layer \(v_2\), they refract towards the normal and as a result no critical refraction is present. As a result, two layers can appear to merge into one layer in the seismic data. This can consequently, cause incorrect imaging as the calculated depth to the next layer can be exaggerated, due to the lower overall velocity of the two combined layers (Mussett & Khan, 2000). In summary, a velocity inversion makes layer thicknesses and interface depths difficult to calculate accurately.
1.3 Ross Ice Shelf and Geological Setting

The RIS is the largest ice shelf in the world. It buttresses the EAIS ice flows from the Transantarctic Mountain (TAM) outlet glaciers, and ice flows from the WAIS ice streams (Figure 1.11; Denton, 2000; Rignot et al. 2008; Muto et al. 2013; Anderson et al. 2002). The RIS is an important interface between the two ice sheets and the ocean, and understanding the ice shelf’s response to past climatic change is important for predicting and making projections of future change. The collapse of the RIS would accelerate global sea level rise (Golledge et al. 2015), so determining the rates at which physical processes occurred in the past during ice shelf and ice sheet retreat can improve our models for future change predictions. It is therefore important that we understand processes, and process interactions, of the ice shelf and surrounding ocean system.
Ice shelves are among the most vulnerable aspects of the Antarctic glaciological system (DePoorter et al. 2013). They are in contact with both the ocean and the atmosphere, both of which are dynamic and changeable (Bromirski & Stephen, 2012). Ice shelves are also the part of the glaciological system that is most reactive to stress, energy, and mass changes (Scambos et al. 2000; Scambos et al. 2009). The potential loss of these buttressing ice shelves makes the WAIS vulnerable to rapid ice loss during climatic warming (DePoorter et al. 2013). A large section of the WAIS lies on bedrock that slopes towards the interior of the ice sheet, and continues below sea level (Muto et al. 2013). Oceanic warming is one of the most effective ways of melting ice (Holland et al. 2008), a process that has been observed at Antarctic ice shelves. The mass loss of ice at the base of ice shelves is increasing, and will continue to do so as the warmer circumpolar deep water is able to reach further underneath the ice shelves (Bintanja et al. 2013). Global sea temperature rise significantly enhances the basal melting rates, and warmer subsurface water can reach underneath the WAIS resulting in the loss of the ice shelves (Muto et al. 2013).
1.3.1 Geological Setting

The RIS, Ross Sea, and part of the continental geology underneath the WAIS are likely underlain by the West Antarctic Rift System (WARS) (Figure 1.12). Little is known about the rift system underneath the RIS or WAIS; however, under the Ross Sea the WARS consists of multiple asymmetric graben structures, separated by basement highs. These sedimentary basins were formed during the Gondwana breakup, and are likely to extend underneath the RIS (Muto et al. 2013). The resulting sedimentary basins are likely filled in with glacimarine sediments (Fitzgerald, 2002; Muto et al. 2013), which were ice entrained debris, melted out, and were deposited on the seafloor. Sediments would be largely detrital and biogenic, with some possible authigenic products (Drewry, 1986).

![Figure 1.12 The West Antarctic Rift System (WARS) encompassing the Ross Sea, Ross Ice Shelf (RIS), and part of West Antarctica (Paulsen & Wilson, 2010).](image)

1.3.2 Past and Future State

Modelling of the Antarctic system, particularly ice shelves, is not well constrained in future climatic conditions (Bindschadler et al. 2013). Increasing surface temperatures...
around the globe, between 0.3 to 4.8°C by the year 2100 and possibly up to 10°C by the year 2300 are a possible response to atmospheric warming (Meinshausen et al. 2011; Rogelj et al. 2012). The reaction of ice sheets to climatic warming is difficult to model as they respond to external forcing on time scales that are different to oceans and the atmosphere. As an example, ice shelf collapse, as a measure of climatic warming, is difficult to quantify (Golledge et al. 2015). Nevertheless, collapse of the ice shelves will increase the flow of ice sheets from behind, which in turn may have a larger contribution to sea level rise (Bintanja et al. 2013). Simulations by Collins et al. (2013) of Antarctic ice sheets’ and ice shelves’ response to four representative concentration pathways (RCPs), show that by the year 2300, all but the lowest RCP will have initiated ice shelf collapse (Figure 1.13).

![Figure 1.13 Multiple climate change models for 2100 CE (left) and 2300 CE (right), and the ice sheet response. The different scenarios for each time frame use the representative concentration pathways (RCPs) used in the Fifth Assessment Report of the Intergovernmental Panel on Climate Change IPCC). The East Antarctic Ice Sheet (EAIS), West Antarctic Ice Sheet (WAIS), and Ross Ice Shelf (RIS) are annotated. Note the disappearance of the RIS at RCP 8.5 in 2100 CE and from 4.5 RCP onwards for 2300 CE. Modified from Golledge et al. (2015).]
Studying the response of the RIS during past periods of climatic change such as those of the Pliocene (5.33 to 2.58 Ma), Pleistocene (2.58 Ma to 11.7 ka) and the Holocene (11.7 ka to recent), may lead to a better understanding of how Antarctica may respond to future warming. During the Pliocene the Ross Ice Shelf and WAIS were unstable and retreated episodically and rapidly under atmospheric CO₂ concentrations were similar to today, and global temperatures similar to those predicted for the year 2100 (Naish et al. 2009), therefore this period could be viewed as an analogue of future warmer state. The retreat timing of the Ross Ice Shelf since the Last Glacial Maximum (LGM) is still poorly constrained in places; however, recent work by McKay et al. (2016) suggests that the retreat of the LGM ice sheet was initiated between 18 and 11 ka. Improved knowledge of how the RIS has responded and changed since the LGM can provide valuable constraints on past, natural rates of change under low CO₂ conditions.

1.4 Previous and Current Work

Three studies provide relevant geological information about previous and current research conducted on the RIS. The Ross Ice Shelf Project (RISP) and Ross Ice Geophysical and Glaciological Survey (RIGGS) were conducted from 1972 to 1978 and produced a grid of data points 55 km apart over the RIS (Bentley, 1984). In 2006 and 2007 the Antarctic Geological Drilling project (ANDRILL) recovered sedimentary drill cores at two sites, AND-1B and AND-2A, coring approximately 1200 m each, with the oldest sediments dating back to 20 Ma (Acton et al. 2008-2009). A current study, the ROSETTA project, is investigating the ice shelf, ocean, and bathymetry beneath the RIS through airborne surveys.

1.4.1 Ross Ice Shelf Project (RISP) and Ross Ice Geophysical and Glaciological Survey (RIGGS)

In 1969 it was suggested that a multidisciplinary, drill core through the RIS would benefit multiple disciplines because it would enable sampling of the ocean and seafloor beneath the ice shelf in addition to the ice shelf itself (Bentley, 1984). The resulting multinational Ross Ice Shelf Project (RISP) included many institutions. To select the drill core location, preliminary studies of the ice thickness and water depth would have to be conducted. It was soon apparent that a comprehensive study of the
RIS would be beneficial. As a result, the Ross Ice Shelf Geophysical and Glaciological Survey (RIGGS) programme began in the austral summer of 1973-1974, and continued for five more seasons, with the last season during the 1977-1978 austral summer. A drill hole location for RISP was selected in the first season. Measurements for RIGGS were conducted on a 55 by 55 km grid, and covered the entire RIS. Multiple geophysical measurements were made at each grid location, including strain rates, gravity, seismic velocities, electrical resistivities, and accumulation rates (Robertson & Bentley, 1990). To determine the ice thickness and water depths, a combination of radio waves and seismic waves were used. The seismic waves do not characteristically display a strong reflection of the ice-water interface, whereas the radio waves are unable to penetrate seawater, and therefore a combination of the two techniques were used. The seafloor depths from the RIGGS survey indicate a trough and saddle type bathymetry near the vicinity of the proposed future hot water drill site (HWDS-2; for which this current study has characterised the seafloor), similar in nature to horst and graben structures observed in the Ross Sea (Figure 1.14; Muto et al. 2013).
Figure 1.14 Map of the Ross Ice Shelf Geophysical and Glaciological Survey (RIGGS) measurement locations and the Ross Ice Shelf Project (RISP) drill core location (J9), on the Ross Ice Shelf (RIS), with the RIS seismic survey location (HWDS-2), Scott Base, and the ANDRILL sites. EAIS = East Antarctic Ice Sheet, WAIS = West Antarctic Ice Sheet. (RIGGS data: Thomas et al. (1984); BEDMAP-2 data: Fretwell et al. (2013))

1.4.2 Antarctic Geological Drilling (ANDRILL)

The Antarctic Geological Drilling (ANDRILL) Programme consisted of two main drill cores (AND-1B on the McMurdo Ice Shelf, and AND-2A in the Southern McMurdo Sound). The aim of the ANDRILL programme was to understand the later Miocene and Pliocene behaviour and variability of the Ross Ice Shelf and the West Antarctic Ice Sheet (WAIS), and how they influence the global climate, ocean circulation, and sea level (Naish et al. 2007a). The McMurdo Ice Shelf (MIS) project recovered a 1258 m long drill core (AND-1B) that provided a record for the Miocene to Pliocene, which showed approximately 60 cycles of interglacial marine, ice proximal, and subglacial
environments (Naish et al. 2007b). At the Southern McMurdo Sound (SMS) drill site a 1138 m sedimentary core was recovered (the AND-2A core) which contained a record of ice sheet variability through the early Miocene (20.2 to 14.5 Ma) (Acton et al. 2008-2009). The core contained a cyclical history of glacial advance and retreat, and varying water depths in this location (Acton et al. 2008-2009). Multiple seismic site characterisation studies were conducted prior to drilling both the AND-1B and AND-2A cores (e.g. Bannister, 1993; Bannister & Naish, 2002; Horgan et al. 2003; Horgan et al. 2005; Henrys et al. 2006; Johnston et al. 2008), and correlation of the drill cores to seismic lines was possible (Figure 1.15; Naish et al. 2007b).

Figure 1.15 Integrated diagram of the AND-1B core that connects seismic reflection data from the McMurdo Ice Shelf (MIS) (a), with core velocity (c) to produce a time-depth conversion (b). This is then also integrated with core densities (d) and lithologies (e), and correlated to seismic stratigraphic units (f) (Naish et al. 2007b).

1.4.3 ROSETTA Project

A current multi-disciplinary and multi-institutional project also conducting research on the RIS, is the ROSETTA project. Aptly named after the historic stone written in three scripts allowing them to be decoded, the ROSETTA project attempts to map three
different Earth systems, the ocean, the ice, and the underlying geology, which will help understand past Antarctic climate history, and predict future responses (Tinto et al. 2014). The ocean system, including mixing in the Ross Sea, underneath the ice shelf, ocean circulation, and tidal processes, are sensitive to geological structures below the ice shelf, as well as the ice shelf itself. In turn, the RIS is sensitive to atmospheric and oceanic changes. The IcePod imaging system is the platform used for the ROSETTA project, and this instrument collects a full dataset of the surface (through LIDAR and surface imagery), and ice thickness and accumulation (through the use of radar and airborne gravity methods). A magnetometer and two gravimeters are also used for identifying the geology and depth of the seafloor below the ice shelf (Turrin, 2015). Figure 1.16 illustrates the current extent of bathymetric knowledge beneath the RIS, with an overlay of bathymetry profiles along conducted survey lines.

![Bathymetric map beneath the Ross Ice Shelf (RIS)](image)

**Figure 1.16** Bathymetric map beneath the Ross Ice Shelf (RIS) based on the low-resolution RIGGS data measurements. Four gravity profiles are overlain, which were collected as part of the IceBridge project, a precursor to ROSETTA. Cross sections are shown for each profile line, where the back lines are BEDMAP2 digital elevation models, red lines are the bathymetry modelled from gravity data, light blue are ice thickness, and dark blue are oceanic water thickness (modified from Tinto et al. 2014).

### 1.5 Aims and Objectives

This study has two main research objectives and one minor objective that will contribute to the New Zealand Antarctic Research Institute (NZARI) larger RIS programme.
1. The first main objective was to **characterise seafloor bathymetry and substrates** in the vicinity of a future hot water drill site that **created an adequate depiction of the subsurface**, and provided information to make informed decisions concerning seafloor coring/sampling locations in following field seasons.

2. The second main objective was to **characterise the roughness of the ice shelf’s basal surface**, which will facilitate the deployment of oceanographic equipment in following field seasons.

3. Due to the lack of high resolution bathymetric data from this locality, data quantity, as well as quality, can be considered a minor objective.

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**Figure 1.17** Map of the Ross Ice Shelf (RIS) with the RIS seismic survey location (HWDS-2) approximately 323 km from Scott Base, and the ANDRILL sites near Scott Base. EAIS = East Antarctic Ice Sheet, WAIS = West Antarctic Ice Sheet. (BEDMAP-2 data: Fretwell et al. (2013)). There is a slight discrepancy between the coastline data (black) and the ice shelf extent (light blue).
Part II - Preliminary Work
2. Assessment of Snow Streamer and Spiked Geophone Acquisition on an Ice Shelf

Prior to selecting the seismic source and receiver package for the RIS seismic survey, a comparative study was conducted between conventional spiked geophone data and snow streamer data collected on the McMurdo Ice Shelf (MIS). This chapter discusses the methods by which two sets of seismic data were collected on the MIS in 2005, as well as main seismic processing steps. The main components of seismic processing, pre-stack processing and velocity analysis, are explained in further detail with examples given. The main motivation for this receiver comparison is to assess the data quality recorded by snow streamer geophones when compared to spiked geophones, considering the potentially poorer ground coupling. The comparison of data quality in the final processed lines was used to constrain the selection of receivers for ongoing experiments on the ice shelves.

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2.1 Survey Design

This receiver comparison uses part of a seismic line (MIS4), collected in 2005 on the MIS, during the austral spring/summer (Figure 2.1). During the collection of conventional geophone data, a snow streamer system was deployed, for a day of coincident acquisition. The geophone data had previously been processed (Johnston et al. 2008; Henrys et al. 2006) and similar data processing methods were used for this study; however, the snow streamer data were not processed, and consequently no comparison between the two was made (Betterly et al. 2007).

![Figure 2.1 Map of the McMurdo Ice Shelf (MIS) seismic lines collected during the austral spring/summers of 2003-2005. This study looks at the comparison of shots on line MIS-4 (purple) where a snow streamer was deployed adjacent to spiked geophones (Henrys et al. 2006).](image)

The MIS seismic survey consisted of five multichannel seismic lines, approximately 43 km in total length, with three lines in the Windless Bight region (33 km in total) and
two lines on the southern MIS (10 km). They were collected over a period of three years as part of site surveying for the ANDRILL sites (e.g. Naish et al. 2007c). The survey aimed to extend coverage of previously collected seismic lines (e.g. collected by Balfour, 2002; Bannister & Naish, 2002; Horgan & Bannister, 2004; Horgan et al. 2003; Horgan et al. 2005; Johnston et al. 2008), as well as establish new lines that intended to extend seismic coverage, and allow for the extension of mapping key stratigraphic horizons.

2.2 Data Acquisition

The following data acquisition information (Figure 2.2) was collected from a field report by Henrys et al. (2006). Initial positioning of the seismic lines was pegged along a global positioning system (GPS) bearing, and pegged. Shot locations were marked every 96 m, and geophone locations marked every 48 m. Each shot was located halfway between two geophones, drilled and loaded with explosives, then tamped with drill shavings and snow. The drill shavings and snow were important for generating high-quality seismic signals. The shots were detonated using an encoder-decoder radio detonation system, which allowed simultaneous detonation and triggering of the seismograph. The shots were fired in batches of 8 to 25.

*Figure 2.2* Vehicle set up for data collection on the McMurdo Ice Shelf (MIS; top image), and snow streamer configuration (bottom image) (Speece et al. 2009).
A total of 96 channels, on sixteen cables with six take-outs each, were deployed from the back of a snow mobile, and were offset from the shot points. The purpose of this was to reduce any noise in the data that may have been induced by physical movement from the shot detonation; it also results in a more even fold distribution. A 96 channel roll switch allowed for maximum roll capability. Data were recorded using two Seistronix RAS-24 portable seismographs and PC operating software, both of which were set up in the back of a Hägglund tracked vehicle. Data were initially recorded in SEG-2 format and later converted to SEG-Y format. Some preliminary pre-stack and post-stack data processing was performed using GLOBE Claritas software v6.3.1.12576 (Ravens, 2001). Table 2.1 contains basic acquisition information used by Henrys et al. (2006) and Johnston et al. (2008).

<table>
<thead>
<tr>
<th>Source Type</th>
<th>Spiked Geophones</th>
<th>Snow Streamer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shot Interval</td>
<td>Pentex PPP primers 3.2-kg explosive with Red cord detonation cord. Electric, 30-m lead seismic detonators formed the primary explosive.</td>
<td>96 m, on line</td>
</tr>
<tr>
<td>Shot Depth</td>
<td></td>
<td>17-18 m</td>
</tr>
<tr>
<td>Number of Live Channels</td>
<td>48</td>
<td>60</td>
</tr>
<tr>
<td>Survey Geometry</td>
<td>Split symmetrical spread (24:24), except for roll off at western end of each line</td>
<td>Off end</td>
</tr>
<tr>
<td>Geophone Type and Spacing</td>
<td>Single string 40 Hz vertical component with 40 cm spikes, 48 m spacing, 2 m offset to south of line</td>
<td>Gimbaled geophones, 30 Hz velocity sensors. Velocity sensors are 360° roll gimbaled, with a 180° pitch tolerance. 25 m spacing.</td>
</tr>
<tr>
<td>Sampling Rate</td>
<td>1 ms</td>
<td></td>
</tr>
<tr>
<td>Record Length</td>
<td>6 s</td>
<td></td>
</tr>
<tr>
<td>Low-Cut Acquisition Filter</td>
<td>10 Hz</td>
<td>None</td>
</tr>
<tr>
<td>High-Cut Acquisition Filter</td>
<td>None</td>
<td></td>
</tr>
</tbody>
</table>

### 2.2.1 Spiked Geophones

The spiked geophones were single 40 Hz vertical-component geophones, placed at an offset of 2 m from the geophone pegs. Forty centimetre spikes were used rather than conventional 10 cm spikes in order to improve coupling in the snow (Figure 2.3).
2.2.2 Snow Streamer

Snow streamers (also called land streamers) are similar to marine streamers with the exception of using gimballed geophones as opposed to hydrophones. The geophones are arranged in an array attached to strapping, and towed behind a vehicle, much like a streamer behind a boat. Snow streamers increase productivity compared to spiked geophones, as well as only requiring a small crew to operate. A challenge of snow streamers is ground coupling in the snow, which is better achieved using spiked geophones (Speece et al. 2009). Marine streamers, by their nature, display excellent coupling with water and can produce very clean seismic images.

The snow streamer used on the MIS, had 60 geophones with 25 m spacings. The geophones were vertically orientated, 360° roll gimballed geophones, with a 180° pitch tolerance, were attached to a snow streamer cable (Figure 2.4). The geophones have 30 Hz velocity sensors, and weigh approximately 1 kg each.
2.3 Data Processing

The seismic processing was done by the author using the GLOBE Claritas software package produced by GNS Science (Ravens, 2001). A general processing sequence was established for the seismic data that was similar for both the spiked geophones and snow streamer. Figure 2.5 outlines the basic processing sequence, and each component is explained in greater detail in subsequent sections. Information pertaining to each processing step was derived from Sheriff (2002), Yilmaz (2008) or Ravens (2001). The above sources will not be referenced again for this section, whereas others will be included.
2.3.1 Pre-Stack Processing (Blue)

2.3.1.1 Data Input

Data input was the first step of seismic processing and required the modification of field file formats, so that they could be read by GLOBE Claritas or other seismic data processing programs. The industry standard for seismic data are SEG-Y files (Barry et al. 1975), but SEG-D or SEG-2 are often used for field files and require conversion to SEG-Y format (Drijkoningen & Verschuur, 2003). By convention, Claritas files are indicated by a .csgy file extension, which is compatible with other SEG-Y files and seismic processing programmes, but with the addition of extended file headers that contain parameters specific to the Claritas software. Prior to initiation of the present study, spiked geophone data were converted to a SEG-Y format, whereas snow streamer data remained in SEG-D (.dat) format files, and were later converted using a script available at the University of Otago.
2.3.1.2 Trace Editing

Trace editing was the second step in data processing, where unwanted traces or shots were removed to eliminate undesirable data elements that can affect outcome quality (Drijkoningen & Verschuur, 2003). Trace editing can also be used to correct any reversed polarities (e.g. geophones that had been wired in reverse). Using Claritas software, the TREMOVE module allowed for the removal of dead, non-live, or all traces (Figure 2.6). For the spiked geophone data, non-live traces were removed. For the snow streamer data, all non-live traces were removed, as well as extraneous traces (channels 61-71 contained no subsurface information), and bad shots (311, 312, and 335, which were two test shots and a dead shot, respectively). The snow streamer data retained 25 of the original 28 shots, recorded also by the spiked geophone array.

![Figure 2.6](image)

**Figure 2.6** Shot 320 from the snow streamer data. No trace editing or static corrections were applied here. The red arrows indicate where a time shift will be applied and traces that need removal.
2.3.1.3 Static Correction

Each shot was manually triggered at the time of the seismic source explosion however as a result, the direct arrivals do not trace back to time zero (0 ms) as they intended to. As a result, time shifts were correctively applied to the snow streamer shot records. Each shot had a different time by which it had to be shifted (Figure 2.6). These were put in a static shift file (.shf) that was used in the Claritas STATIC module. The spiked geophone data did not need an initial time shift as all the direct arrivals traced back to 0 ms.

2.3.1.4 Adding Geometry

Adding geometrical information to the seismic data and sorting it into a common midpoint gather (CMP) was an important step of data processing. Adding traces together increased the signal-to-noise ratio of seismic data, therefore improving data quality. In this step of the processing sequence, field geometry was incorporated into the headers of the SEG-Y files. Data in the field are sequentially, where all receivers record traces (measurements at one receiver location) for each shot, numbered in either increasing or decreasing offset away from the source. From here, the source was moved to the next shot point, and the recording process was repeated. Data were ordered by common-shot gather (Drijkoningen & Verschuur, 2003); however, common-shot gathers are not the best way of organising seismic data for further processing. Following data ordering, the data were sorted based on a common midpoint gather (CMP) (or equivalently common depth point gather (CDP); Figure 2.7). Once the traces had been sorted into CMPs or CDPs, they were further ordered by increasing or decreasing offset; even though the records have a range of different offsets, they all contained information relating to the same subsurface points.
Figure 2.7 A common shot gather (a) and common midpoint gather (b) (Drijkoningen & Verschuur, 2003).

Shot receiver locations, shot spacing, and shot coordinates were entered into the system and are used to calculate CDP locations. The accuracy of CDP sorting is important, as incorrect geometrical information can lead to errors such as false positives in the final processed data (Figure 2.8).

Figure 2.8 Incorrect geometry from snow streamer data (a) compared to correct geometry from spiked geophones (b) of the MIS4 line. The snow streamer data appeared to have stratigraphy present on the left hand side of the section, however, during velocity analysis a large element of this was shown to be artefacts of incorrect geometry.

The spiked geophone data had land geometry applied prior to the start of this study. For the snow streamer it was possible to apply either land or marine geometry
conventions as the snow streamer configuration (an off-end array) was similar to the configuration of a marine streamer. When setting up marine geometry in Claritas, a general streamer file (.str) that contained streamer geometry information in relation to the vessel was necessary, as well as separate shot files (.sht) that contained global positioning system (GPS) coordinates for each line. Setting up seismic geometry in Claritas created a formatted geometry (.geom) file that was used to update the headers of the pre-stack data using the ADDGEOM module in Claritas. After the geometry headers were added to the data, the DISCSORT module read the SEG-Y file with geometry information, and output the sorted traces. Figure 2.9 illustrates the areal fold distribution of CDP-sorted spiked geophone records plotted with a New Zealand Transverse Mercator (NZTM) projection.

Figure 2.9 Areal CDP fold of the spiked geophone line, part of MIS4, which shows the highest fold (8) in the centre of the line, with lower fold at the ends. Northings and Eastings as indicated are in metres with an NZTM projection.
2.3.1.5 Bandpass Filtering

Raw seismic traces contained both the sought-after subsurface signal plus noise. If possible, the signal to noise ratio should be increased. One way this was achieved was through bandpass filtering. A bandpass filter was defined by four corner frequencies within a certain range that rejected (or filtered out) data outside that range. Examples of bandpass filters include the Ormsby and Butterworth filters (Figure 2.10). The bandpass routine linearly tapered between the first and last pairs of frequencies, and produced a trapezoidal filter.

![Zero-Phase Ormsby Filter](image)

![Zero-Phase Butterworth Filter](image)

*Figure 2.10* Ormsby and Butterworth bandpass filters. The Ormsby filter (red) has sharper corners compared to the Butterworth filter (blue), where the corner frequencies (indicated by the black dots and grey dashed lines) are much smoother. Modified from Hess Corporation (2015).

The FDFILT module in Claritas is an Ormsby filter and BUTTERFILT module a Butterworth filter. Both modules in Claritas used a bandpass filter with user defined frequencies (22, 45, 140, and 210 Hz) outside which data were rejected. Figure 2.11 shows the effect of no filter, a Butterworth filter, and an Ormsby filter.
Figure 2.11 No filter (a), a Butterworth filter (b), and an Ormsby filter with the same corner frequencies (22, 45, 140, and 210 Hz). The bandpass filters had minimal effect on the subsurface data, and was most noticeable in the surface waves.

2.3.1.6 Gain Recovery

Gain Recovery in seismic data adjusted for energy decay between the source and the receiver. The greater energy close to the source resulted in higher amplitudes at the top of traces which can overshadow lower amplitudes at depth, and the traces needed to be equalised (Drijkoningen & Verschuur, 2003). There are three main reasons why the energy decays with time; geometrical spreading, energy absorption in firn and glacial ice, and transmission loss at interface boundaries (e.g. the ice-water interface). The wavefront geometrically spread through the medium, and the amplitude decayed as it did so. This wavefront spreading was a result of a travel time and average spreading velocity, and was corrected using a spreading function. The assumption was made for the spreading that it did so in three-dimensional space. An exponential gain
function was used to correct absorption, where average absorption coefficients of the rocks were related to the correcting function (Drijkoningen & Verschuur, 2003). Lastly, transmission loss occurred at the interface between layers in the subsurface, but was generally corrected by an exponential function for absorption.

Multiple modules in Claritas allowed for gain recovery and correction. Automatic Gain Control (AGC), trace balancing (BALANCE), and a spherical divergence operator (SPHDIV) were trialled for the MIS data. The AGC module is an operator that scaled a trace sample by sample, by normalising the amplitude of a sample by the average amplitude of a specified window (250 ms) centred on that window (Figure 2.12). The overall result of this operator was that the average amplitude for a trace was constant. The BALANCE module uniformly scaled a trace, and was beneficial for balancing the traces of a gather or stack before 2D processing (e.g. f-k filtering). The BALANCE module affected the horizontal trace balance (i.e. space), instead of vertical (i.e. time), which was done by ACG and SPHDIV modules. Lastly, the SPHDIV module compensated for seismic wave amplitude attenuation caused by geometrical wavefront spreading, and energy dissipation. Multiplying the traces by a scaling function removed this attenuation (Drijkoningen & Verschuur, 2003).
2.3.2 Velocity Analysis (Red)

Seismic wave velocities provide the means to link time and depth, and therefore accurate velocity analysis is arguably the most important parameter in seismic processing, and was necessary to carry out a time-to-depth conversion (Drijkoningen & Verschuur, 2003). Consequently, a large component of seismic processing was devoted to velocity analysis. Data need to be pre-processed before detailed velocity analyses can take place. This can involve deconvolution, CMP sorting, muting, filtering, producing rawstacks and even pre-stack migration in some cases (Hess Corporation, 2015).
2.3.2.1 Rawstack

To increase the signal-to-noise ratio of a seismic section, data were stacked. In the raw data, only strong reflections are seen, and weaker ones are often lost among noise (Drijkoningen & Verschuur, 2003). The rawstack is the term used for the initial stacking of CDP traces that have been added together prior to accurate velocity analysis. By creating and using a velocity model, normal moveout (NMO) corrected traces were stacked into a CDP gather, which created one output trace for each CDP with an increased signal-to-noise ratio. This was an iterative process where, as the velocity model was updated and improved, the quality of the stack increased (Figure 2.13). File-size reduction of data occurs when a section is stacked, and this can be beneficial for further stages in seismic processing, such as migration, which can be time consuming to perform, and are sometimes better applied post-stack than pre-stack (Drijkoningen & Verschuur, 2003). The STACK module in Claritas created a stack of traces from a set of CDP gathers.

![Figure 2.13](image)

**Figure 2.13** Comparison of a rawstack made with the initial velocity model (a) to the stack created using the final velocity model (b). Note the correct stacking of seismic layers around 1200 ms. No AGC or filtering was applied to these stacks.
2.3.2.2 Normal Moveout Correction

NMO describes the effect that the distance between a seismic source and receiver, the offset, has on the return time of energy. This additional return time makes flat reflections appear approximately hyperbolic in shape (Figure 2.14). Velocity analysis enabled the hyperbolic moveout of seismic reflections to be corrected for prior to stacking. The hyperbolic shape of reflections was one of the principal ways to determine whether a seemingly correct reflection was actually a seismic artefact, or the result of refractions, multiples, and diffractions. When velocity analysis was done correctly, real horizons appeared flat, refractions as inverted curves, and multiples and diffractions remained slightly curved.

Figure 2.14  Schematic diagrams showing the ray path in a single layer (a) and a two layer (b) scenario. The seismic response is seen on the right, and shows the hyperbolic shape of the seismic layers (a one-layer case will be exactly hyperbolic, whereas a two-layer case will be approximately hyperbolic due to refraction at the boundary).
In Claritas, the Claritas Velocity Analysis (CVA) tool allowed for accurate velocity analysis and velocity file creation. This tool required a rawstack and CDP sorted .csgy file as inputs, thus the creation of an initial rawstack with an initial .nmo file was required. The velocities were picked along discrete intervals and the model interpolated between the picked velocities. The quantity of velocity values picked depended on the complexity of the geology, but there were sufficient picks to define the geological structures. Velocity picking was a partially interpretive process, as the interpretation and preferences of the picker determine the methods used and the velocities picked. The two main sub-tools used within CVA were the constant velocity stack (CVS) and semblance modules. The CVS used multiple adjacent CDPs to create a mini-stack, and NMO corrected the CDPs at a range of velocities. The mini-panels can be viewed in succession and displayed which velocity shows the greatest continuity or highest amplitude for a horizon. Data with a velocity too high or too low becomes obvious using this technique (Figure 2.15).
Figure 2.15 Spiked geophone data with constant velocity stacks that have been NMO corrected for three different velocity values. The velocity is too low at 1300 m/s, and the seafloor reflection does not stack in properly. The correct velocity at the seafloor is 1500 m/s, and the horizon is coherent. At 1700 m/s the NMO hyperbolae are not corrected fully, and can still be seen just below the seafloor reflection.
The semblance display was a measure of the coherency. Semblance is more tuned to the seismic signal rather than noise interference, lateral amplitude variation, and spatial aliasing than the CVS display (Hess Corporation, 2015). A high semblance value was shown in red and low semblance in blue, and thus a high semblance point was relatively easy to distinguish from the rest of the data (Figure 2.16). Multiples were easily distinguished in a semblance display, and appeared as high amplitude areas, but at velocities that were similar to their primary arrivals.

![Figure 2.16 Semblance window showing the coherence between where a hyperbola best fits a high amplitude seismic horizon. The x-axis shows NMO velocities, and the y-axis two-way travel time.](image)

### 2.4 Results

No geological interpretation was made on either the spiked geophone or snow streamer data. The aspects compared were the frequency spectra of both receivers, a visual shot comparison, and a rawstack comparison.
2.4.1 Amplitude Spectrum Graph

The amplitude spectrum graph allows for a comparison of the frequency content of the seismic data sets. The spectrum shows the relative amplitude across a band of frequencies by calculating a Fourier transform of the input seismic data. The spiked geophone data had a field filter applied, as observed in Figure 2.17, at a clear frequency range (5 to 135 Hz), which had a higher average amplitude.

![Amplitude Spectrum Graph](image)

*Figure 2.17* Amplitude spectrum graphs of shot record 102 for the spiked geophone data (blue) versus shot record 313 for the snow streamer data (red). The spiked geophone data had a field filter applied, whereas the snow streamer data did not. However, both receiver types recorded similar frequency spectra, at similar amplitude levels.

2.4.2 Visual Shot Comparison

A visual shot comparison of shot record 313 (snow streamer) and shot record 102 (spiked geophones) illustrated the similar quality of the data. Spiked geophone channels 1 to 24 were comparable to snow streamer channels 16 to 60. These were the result of different trace spacings, which displayed similar seismic features (Figure 2.18). Both shot records displayed direct arrivals and the seismic response of the MIS in approximately the first 1000 ms of the shot record. The seafloor reflection was seen at zero offset at 1250 ms (i.e. at the source location; approximately channel 24 for the spiked geophone data and channel 60 for the snow streamer data). Internal reflections were observed on both shot records between 1330 and 1750 ms, where further the seafloor multiple was seen at approximately 2400 ms. The quality of both shot records was virtually identical.
Figure 2.18 Visual shot comparison of the spiked geophones and the snow streamer. Channels 1 to 24 on the spiked geophone shot record covered the same amount of ground as channels 16 to 60 on the snow streamer, and display similar characteristics. The direct arrivals, seafloor, and seafloor multiples are highlighted in orange.

2.3.4 Rawstack

The geometry of the snow streamer data was incorrect; nevertheless, the data quality could still be distinguished. Visually, both data sets were of a similar quality. Figure 2.19 displays the seafloor in the rawstack for both the snow streamer and spiked geophone data, at approximately 1250 ms.
Figure 2.19 Rawstacks of the snow streamer (a) and spiked geophone data (b). The geometry of the snow streamer data was incorrect; however the data quality could still be distinguished. Both data sets visually were of a similar high quality.

2.5 Conclusions

The data quality of the snow streamer system were comparable to that of the spiked geophones and demonstrated multiple advantages over the spiked geophones:

1. It was a system that allowed for greater quantities of data to be collected compared to spiked geophones in the same time frame.
2. It required smaller crew numbers to man, and less manual effort to operate in contrast to the previously used method of pushing spiked geophones into the ground, removing the geophones during roll-over, and replacing them.
3. The snow streamer was attached to a vehicle or the seismic source, and towed behind the recording system which allowed for faster data collection.
4. Closer spacing of streamer geophones, which record data at a higher resolution.
The two disadvantage of this system are:

1. Coupling to the ground was likely better with the spiked geophones; however, snow streamer data are still of high quality.
2. Offset length of the streamer cable was shorter than the spiked geophones, which could hinder data collection at greater depths.

Nonetheless, the advantages of the snow streamer system outweigh the disadvantages, and may help address the third objective for this study (see section 1.5). These qualities, as well as the high quality of the data collected, make the snow streamer the preferred choice, was selected to carry out all further experiments and discussed later in this study.
3. Survey Design: Ross Ice Shelf

Due to the remoteness and difficulty of reaching the Ross Ice Shelf (RIS), it was important to thoughtfully design the seismic survey in order to produce accurate and scientifically relevant results. The following chapter discusses many aspects involved in the design of the seismic survey, and explores each aspect before seismic field work was undertaken in Antarctica.

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3.1 Survey Design

Evans (1997) described technical considerations that should be made before setting up a seismic survey in an ice environment, the station spacing, energy source, and crew numbers. Figure 3.1 shows the different aspects of the seismic survey that were taken into consideration, and are discussed in later sections.

![Figure 3.1](image)

Figure 3.1 Seismic data acquisition set up, with various components of the system labelled, and discussed in this chapter. The receivers are represented by green triangles and the seismic source by a star.
3.1.1 Cable Length
In order to produce high quality data, the far-offset receivers were a similar distance away from the seismic source as the target horizons are below the surface. The distance of the near-offset receivers away from the source should be determined by how much noise is generated by the source. If the offsets are too great, more of the seismic energy will be recorded as refractions rather than reflections. If offsets are too small, velocity analysis becomes difficult and less accurate. In the present study, the snow streamer had a cable length of approximately 300 m between channels 1 and 96, therefore the far-offset was approximately 300 m away from the near-offset.

3.1.2 Station Spacing
In order to avoid spatial aliasing in the data, receivers should be positioned closely together. When collecting seismic data on ice above water, flexural waves can be highly dispersive, with low frequencies at low velocities and high frequencies at high velocities (Henley, 2005). A receiver spacing of 3 to 5 m has been shown to effectively attenuate this wave (Henley, 2005). The snow streamer used for the RIS had a receiver spacing of 3.125 m, which was able to sufficiently sample the data without introducing spatial aliasing.

3.1.3 Channel Quantity
Channel quantity is often limited by the capacity of the recording equipment. The snow streamer used for the RIS study had 96 channels. The total length of the snow streamer was close to 300 m, which was sufficient to image the primary outcome (the sea floor). There was a possibility that the geophone spacing would allow us to image the basal interface of the RIS.

3.1.4 Geophone Array
The geophones used were gimballed geophones linked in a linear array. Gimballed geophones are known to be more than adequate for towed streamer configurations, as they are self-levelling. Due to the nature of the snow streamer, this array was an off-end spread, where all the receivers were on one side of the shot point. This was in
contrast to the more common split spread often used in land seismic data, where the source location is in the middle of the geophone array.

3.1.5 Detector Frequency
The natural frequency of a geophone becomes important when disproportionate low-frequency ground-roll is expected to reduce the signal of reflected signals. We were limited to the availability of geophones that came with the snow streamer, which had a natural frequency of 30 Hz. As a result at frequencies below 30 Hz, there is a drop off in the system’s ability to record the signal.

3.1.6 Fold
The fold of a seismic reflection profile corresponds to the number of traces recorded for each subsurface bin in the resulting image. This was defined as:

\[
Fold = \frac{\text{Number of channels}}{2 \times \left( \frac{\text{Shot interval}}{\text{Group interval}} \right)}
\]

Equation 3.1 Fold calculation to determine shot spacing (Mussett & Khan, 2000)

Seismic reflection fold needs to be considered to accurately image the subsurface. A higher fold is preferred, as it increases the signal-to-noise ratio and produces cleaner reflections. Using the above equation for the RIS active seismic survey, a 12.5 m shot spacing with a 96 channel and 3.125 m spacing streamer, results in a fold of 12, whereas a 25 m shot spacing would provide a fold of 6. Through the use of a thumper system, it is likely we could create a shot point every 1 to 2 minutes, thus a shot spacing of 25 m would effectively double the ground we could cover. The decision needed to be made as to whether it was preferable to have a larger quantity of data at a lower quality, or alternatively, to survey a smaller area but at a higher resolution. This was a decision which was made on the ice, where both 12.5 m and 25 m shot spacings were trialled.
3.1.7 Line Length

Line length becomes an important consideration as data were collected. The tail ends of a seismic line have lower fold, and as a result are of lower quality than the rest of the data with a maximum fold. If during data acquisition, the feature of interest is imaged near the tail end of a line, the line will be extended so that a maximum fold can be achieved.

3.1.8 Energy Source

The seismic source can affect both the signal-to-noise ratio and resolution of the acquired data. The weight drop seismic source (WDSS) was custom designed for the purpose of data collection on the RIS (Appendix II), and was selected using the following five criteria:

1. Depth Penetration: In order to image the seafloor (the target horizon), the WDSS was determined to be a source with sufficient energy, approximately 3140 J (Appendix II).
2. Signal-to-noise ratio: Differing geological areas highlight a variety of different problems with regard to data noise and hence may dictate the source selection. As an example, the RIS ground roll can be a large component of the returning signal from the ice, which might contribute to increased noise generation (Beaudoin et al. 1992). Currently, it is not known how well the WDSS will behave in this environment.
3. Repeatability: The often used technique of a sledge hammer on a steel plate can result in increased variability and reduced repeatability, due to the influence of human error. In contrast, WDSS displays a consistent repeatable waveform which is less influenced by human error, and therefore may provide higher quality data.
4. Environment: Ice environments often display a unique set of characteristics which need to be first identified when selecting an energy source. The firn layer has been seen to attenuate much of the seismic energy created by an energy source, therefore when recording on the firn layer it is preferable to choose the source with the greatest energy output.
5. Portability: As the WDSS is smaller than a vibroseis, its use can be an advantage because it can be towed behind a vehicle together with the streamer.

In contrast to the above considerations, the limitations of the WDSS include a double bounce from the weight drop, which could be observed in test seismic data. Nevertheless, the WDSS acts as a superior energy source compared to other techniques, and therefore was chosen as the most appropriate and effective seismic source to fulfil our primary goal of imaging the sea floor.

3.1.9 Geographic Positioning

Accurate geographical positioning of the receiver array and shot points is necessary for two important reasons. Firstly, data processing required accurate relative source and receiver positions for geometry application, so that the information can later be mapped. Secondly, if it is required to tie multiple seismic lines together, it becomes very important that their relative positions are accurate, otherwise errors are introduced at both the processing and interpretation steps.

The combination of the multiple satellite systems allowed for greater global coverage than the use of the GPS system alone (Capra & Dietrich, 2008). For this reason, the positions of the receiver array and shot points were recorded using a Global Navigation Satellite System (GNSS) consisting of American Global Positioning System (GPS) satellite data, the Russian GLObal NAvigation Satellite System (GLONASS) data, and other systems from Japan and China. Shot points were located in real-time using live GNSS positioning along a pre-determined survey line.

3.1.10 Statics

Static are defined as reflections from seismic events that arrive at different times in each trace requiring time shifts (Drijkoningen & Verschuur, 2003; see section 2.3.1.3). These are typically introduced into seismic data as a shift in time caused by topography or shallow sub-surface velocity anomalies. Static shifts from topographic variations were unlikely in the present study, due to the lack of topographical relief on the RIS.
Where a firn layer of variable thickness is present, minimal significant velocity anomalies within the shelf may require minor ‘weathering’ static corrections.

3.1.1 Crew numbers
Crew numbers are often restricted and limited due to transportation, support costs, and space availability. A crew of three was required for the data collection system used in the present study. One crew member was responsible for the driving of the Hägglund vehicle and positioning of the vehicle for shot points. A second crew member (the observer) was positioned in the back cabin of the Hägglund (known as the ‘dog box’), and were responsible for operating the recording gear, archiving the seismic data during collection, and triggering the WDSS. The final member was positioned outside by the WDSS, and were responsible for troubleshooting, ensuring that the WDSS and streamer were in the correct positions, ad-hoc checking of geophones, and undertaking all other streamer maintenance requirements. In order to increase efficiency and flexibility of the small team, all crew members were trained in all roles.

3.1.12 Resolution
Through the employment of optimal techniques, both high lateral and vertical resolution can be achieved. A higher lateral resolution requires closer shot and/or receiver spacings. The drawback of closer shot and/or receiver spacing is that less distance will be covered in the same amount of time. A higher vertical resolution requires a higher sampling rate, however can be limited by the greater data file size associated with increased sampling. This can generally be overcome by using an increased data storage system. The trade-off between resolution and quantity/file size is important but could be decided in the field.

3.1.13 Numbering Convention
In general, lines were numbered chronologically from 01 as straight line segments. When the acquisition system turned a corner, the new line was named. The first shot point used in the present study was 1001 for line 01, 2001 for line 02, etc. and increased in increments of one for each shot (e.g. 1001, 1002, 1003).
4. Synthetic Shot Record Modelling

Prior to field data acquisition, a series of synthetic shot records were generated to test the effect of ice thickness variations and the presence of a firn layer. This was done to indicate how shot records may look in different ice thickness conditions. Ice is known to be a complex medium that can intrinsically effect seismic wave propagation and recording. Through the use of synthetic modelling to account for variables such as reflections, refractions and diffractions at ice boundary interfaces, it is possible to obtain a more accurate and true representation of then underlying geology. The use of synthetic modelling as a technique is still in its infancy and remains untested in an ice environment, and therefore careful analysis and thought was required before interpreting data produced.

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4.1 Synthetic Shot Records

Synthetic shot records and modelling of the seismic data can lead to a better understanding of geology when applied to real data. Using synthetic models, multiple earth models were tested and ascertain how closely the synthetic records resemble real data after collection. The synthetic shot records can be used to determine how multiple reflections within the ice shelf interfere with seafloor and sub-seafloor reflections. The earth models were created using P- and S-wave velocities and densities of various media. Together with waveform modelling or quick ray-tracing, earth models produced synthetic shot gathers that can be further processed and viewed similar to real data (Hodgetts & Howell, 2000). A description of the seismic processing job flow used in Claritas software is shown below in Table 4.1.

<table>
<thead>
<tr>
<th>Function</th>
<th>Module</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Job Control</td>
<td>SEISJOB</td>
<td>Defines project name, line ID, line-specific parameters.</td>
</tr>
<tr>
<td>Synthetic Modelling</td>
<td>SYNSHOT</td>
<td>Modelling of synthetic shots for horizontal layers</td>
</tr>
<tr>
<td>Trace Removal</td>
<td>TREMOVE</td>
<td>Removes specified traces from the processing flow</td>
</tr>
<tr>
<td>Data Output</td>
<td>DISCWRITE</td>
<td>Writes seismic traces to Claritas SEG-Y disc file</td>
</tr>
</tbody>
</table>

4.2 Ice Thickness Variation

Ice shelf thickness was the first variable modelled using the synthetic shot records to determine what effect it had on synthetic shots records, and at what depth the ice was too thick to image the seafloor properly. Ice thicknesses, ranging from 100 to 700 m, with the seafloor consistently at a depth of 700 m (Bentley, 1984), were modelled at 50 m increments. Figure 4.1 shows the basic earth model used to create the synthetic shots with P-wave velocity, S-wave velocity, density, and depth parameters for each layer (for density and velocity values, see Figures 1.8 and 1.9). The only value changed to create the different shot models was the depth at which the ice-water interface occurred.
Figure 4.1 Main earth model used for generating the synthetic shots with varying ice thickness. The only value altered was the thickness of the ice, the seafloor depth was constant for each scenario.

Using this earth model, synthetic shot records were generated to examine the effect of ice thickness on seismic response. In Figure 4.2, an average thickness of 350 m of ice and 350 m of water was used. The depth to interfaces was calculated using the velocity equation:

$$t = \frac{2d}{v}$$

Equation 4.1 Velocity equation rearranged to solve for one-way travel time.

where $t$ is the travel-time (ms), $d$ the depth (m), and $v$ the velocity (m/s). The synthetic shot record revealed the ice-water and water-seafloor interface for the P-wave
velocity at 200 ms and 700 ms, respectively, as well as in multiples of both interfaces. Figures 4.3 and 4.4 show synthetic shot records and earth models for both 100 m and 700 m ice thickness, respectively.

Figure 4.2 Synthetic shot model of 350 m ice thickness, and ocean floor depth of 700 m. The ice-water and seafloor interfaces are clearly observed, as well as the multiples. The interfaces and multiples are also defined by whether they are thought to be a P-wave velocity ($V_p$) or an S-wave velocity ($V_s$) feature.
Figure 4.3 Synthetic shot model of 100 m ice thickness, and ocean floor depth of 700 m. The ice-water and seafloor interfaces are difficult to distinguish, and there are interferences with multiples.
Figure 4.4 Synthetic shot model of grounded ice, at an ocean floor depth of 700 m. Modelling this ice thickness gives an indication of the type of reflections that might be expected. The ice-seafloor interface is seen at 400 ms, and a likely multiple at 600 ms.
4.2.1 Reflection, Transmission, and Polarity of Seismic Waves

Seismic waves can be both reflected and transmitted at an interface resulting in an abrupt change in seismic velocity. The reflection coefficient \( R \) is a ratio of the reflected and incident amplitudes, and the transmission coefficient \( T \) is a ratio of transmission and incident amplitudes; both of which can be used to determine the strength of a reflection in a seismic shot record (Mussett & Khan, 2000). The coefficients are dependent on rock densities \( \rho \), and P-wave velocities \( \nu \), which when combined produce the acoustic impedance of a layer \( \rho \nu \). The equations for each coefficient are as follows:

\[
R = \frac{\rho_2 \nu_2 - \rho_1 \nu_1}{\rho_2 \nu_2 + \rho_1 \nu_1}
\]

\textit{Equation 4.2a} Reflection coefficient equation (Mussett & Khan, 2000).

\[
T = \frac{2\rho_1 \nu_1}{\rho_2 \nu_2 + \rho_1 \nu_1}
\]

\textit{Equation 4.2b} Transmission coefficient equation (Mussett & Khan, 2000).

It has been demonstrated that the larger the contrast between two layers, the stronger the reflection will be (Mussett & Khan, 2000). When seismic waves travel from a high velocity layer to a low velocity layer, the reflection and thus acoustic impedance will be a negative value. The negative value results in the polarity of a seismic trace being reversed, as shown in Figure 4.5. Equations 4.2a and 4.2b are limited by the assumption that the seismic wave is perpendicular to the interface. When the seismic wave is oblique, calculation of the coefficients are more difficult, where mode conversion occurs between P- and S-waves, and results in different \( R \) and \( T \) values governed by Zoeppritz equations (Yilmaz, 2008). Assuming that all energy is either reflected or transmitted, then ratios of reflected and transmitted energies are described through \( R^2 \) and \( T^2 \), respectively, where \( R^2 + T^2 = 1 \) (Musset & Khan, 2000).

Equations 4.2a and 4.2b can be used to calculate the amount of energy that is reflected and transmitted at interfaces of the ice shelf, such as the ice-water interface on the modelled shot records. In this example, the ice was assumed to have a density
(ρ₁) of 900 kg/m³, a P-wave velocity (v₁) of 3400 m/s (for density and velocity values, see Figures 1.8 and 1.9). Further, water had a density (ρ₂) of 1000 kg/m³, and a P-wave velocity v₂) of 1500 m/s. Inputting these values in the reflection coefficient (R) equation, the reflection coefficient at the basal ice-water interface was -0.34, and therefore had a negative acoustic impedance (Figure 4.1.1). Following on, the ratio of the reflected and transmitted energies was calculated so that approximately 12% of the energy was expected to be reflected at the interface between the ice shelf and the water column, and 88% transmitted.

Both Figures 4.2 and 4.3 demonstrated the polarity of the ice-water interface was the reverse of that of the water-seafloor interface. In Figure 4.2 at the ice-water interface, this could be seen as the negative response of the wavelet first (red) and the positive second (black). At the seafloor the positive response of the wavelet was seen first, and the negative second, and a change in acoustic impedance from a medium with a higher P-wave velocity (i.e. glacial ice) to one with a lower P-wave velocity (i.e. oceanic water), resulted in a seismic response with a reversed polarity (Mussett & Khan, 2000).

4.3 Firn Layer

A firn layer was present on the RIS in the area where the seismic study was conducted (section 1.3.1), and was taken into account when creating synthetic shot records. The values used in the following earth model were taken from Thiel and Ostenso (1961). The modelled firn layer is 50 m thick, which was reasonably expected in this area (section 1.3.1) and exponentially increases in P-wave velocity, S-wave velocity, and
density. From 50 to 350 m, normal glacial ice is assumed, underlain by water with a thickness of 350 m (Figure 4.6). This earth model produces a synthetic shot record that is unlike those that have no firn layer. The direct arrivals are different, and the seafloor is difficult to distinguish with the firn present, possibly due to the energy absorption in firn.
Figure 4.6 Earth model used to create a synthetic shot with a 50 m thick firn layer and 300 m of ice present underneath, and exponentially increasing velocities and densities seen on the earth model. The synthetic shot record (right) shows strong direct arrivals that mostly obscure the other seismic data. Firn values used from Thiel and Ostenso (1960).
4.4 Limitations

4.4.1 Ice Thickness

The synthetic shot records indicated that for ice thicknesses of greater than 500 m, the multiple from the ice-water interface interferes with the seafloor reflection, and partially obscures the seismic signal (Figure 4.7). Any ice thickness less than 500 m was not problematic for imaging the seafloor, as the ice has been previously imaged at approximately 350 m thick in this location (Bentley, 1984).

Figure 4.7 Synthetic shot record of 500 m ice thickness, and ocean floor depth of 700 m. The ice-water multiple and seafloor interfaces are recorded at the same time, and the multiple interferes with the real data.

4.4.2 Software

The module used in Claritas software had multiple problems, resulting in some unexplained artefacts found in the data. The user defined number of channels (1164)
had to be considerably larger than the actual number of channels (96), otherwise the sides of the synthetic shot record would result in reverberations within the record (Figure 4.8). This could be avoided by adding 1068 extra channels for the modelling, then removing them with the TREMOVE module in Claritas.

*Figure 4.8* Synthetic shot record with the earth model of 350 m ice thickness. For this shot record only 96 channels were used as an input. The boundaries cause major reverberations of seismic energy within the shot record that obscures any real reflections.
Part III - Ross Ice Shelf Data
5. Seismic Data Acquisition and Data Processing: Ross Ice Shelf

This chapter contains data acquisition methods and parameters used in the processing of the RIS seismic data. The processing methods within this chapter are additional to those found in chapter 2. The following sections discuss four post-stack processing methods, f-k filtering, migration, bandpass filtering, and coherency filtering, and their effect on the RIS seismic data.

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5.1 Data Acquisition

Seismic surveying on the Ross Ice Shelf (RIS) during November 2015 (Antarctica New Zealand Event K061) represents the first phase of hot water drill site (HWDS-2) surveying in the area. Four lines (totaling 45.8 km) of 96-channel seismic data were collected in November 2015 on the RIS, 323 km south of Ross Island (Figure 5.1). They were collected over a nine-day period, with one day of no data collection. The data were collected using a snow streamer, and a WDSS towed by a Hägglund (Appendix II). The seismic lines were collected to image the seafloor bathymetry and associated sediments. As well as the four main lines, eight smaller supergathers were collected at the start and end of each line. The supergatheres were not processed as part of this project.

Figure 5.1 Map of the four main seismic lines (black) and the supergatheres (red), located on an insert map of the Ross Ice Shelf (RIS; red). Arrows indicate direction of acquisition.
5.1.1 Main Lines

The four main seismic lines were collected as an irregular quadrilateral (Figure 5.1), the main purpose of which was to image the seafloor and sediments beneath the RIS in this location. The initial purpose of Line 01 was as a reconnaissance line to select the field camp location. After basic seismic processing, this line proved that this area was adequate as a campsite location, and Line 01 was extended to a total length of 11.6 km. From the east end of Line 01 a perpendicular line, 02, was collected for a total of 14.5 km. Line 03 is the longest line at 17.4 km, at 45° to Line 02 and heads westward towards the campsite. Line 04 is the shortest line, at 2.3 km, and joins the west ends of lines 03 and 01, shot in a northerly direction. Table 5.1 contains basic line information. The seismic lines intersect with data collection locations of other members of the RIS research team, and some of these data were used to make processing decisions (e.g. ice thickness for velocity modelling).

<table>
<thead>
<tr>
<th>Line</th>
<th>First SP</th>
<th>Last SP</th>
<th>Number of shots</th>
<th>First CDP</th>
<th>Last CDP</th>
<th>Length</th>
<th>Direction</th>
</tr>
</thead>
<tbody>
<tr>
<td>L01</td>
<td>1001</td>
<td>1462</td>
<td>462</td>
<td>100</td>
<td>3880</td>
<td>11.6 km</td>
<td>NE</td>
</tr>
<tr>
<td>L02</td>
<td>2001</td>
<td>2579</td>
<td>579</td>
<td>100</td>
<td>4772</td>
<td>14.5 km</td>
<td>SW</td>
</tr>
<tr>
<td>L03</td>
<td>3001</td>
<td>3696</td>
<td>696</td>
<td>100</td>
<td>5704</td>
<td>17.4 km</td>
<td>W</td>
</tr>
<tr>
<td>L04</td>
<td>4001</td>
<td>4093</td>
<td>93</td>
<td>100</td>
<td>880</td>
<td>2.3 km</td>
<td>N</td>
</tr>
</tbody>
</table>

Data were acquired using a geophone snow streamer from the Alfred Wegner Institute in Germany, and a weight drop seismic source designed and built at the University of Otago (Figure 5.2). The snow streamer had 96 channels, with channel one furthest from the source, channel 96 the closest, and channel spacings of 3.125 m. The shot interval was 25 m, and data were recorded using four Geometrics 24-channel Geodes (table 5.2).
5.1.2 Supergathers

At the start and end of each line a short supergather was collected (with the exception of line 01, which only has a supergather at the end of the line). These were done with the geophones left stationary, and the source moving up at 0.39 m increments, to a total of 3.125 m to enable the assembly of a 768 channel record. This record allows for a more detailed study of ice characteristics. These will not be processed as part of this study; however, data acquisition information is provided as a record in relation to the main seismic lines. Table 5.3 shows the different acquisition information for the supergathers.
Table 5.3 Supergather data acquisition information.

<table>
<thead>
<tr>
<th>Line</th>
<th>First SP</th>
<th>Last SP</th>
<th>Number of shots</th>
<th>Equivalent SP on Main Lines</th>
<th>Length</th>
<th>Direction</th>
</tr>
</thead>
<tbody>
<tr>
<td>L101</td>
<td>1001</td>
<td>1009</td>
<td>9</td>
<td>1462</td>
<td>3.125 m</td>
<td>NE</td>
</tr>
<tr>
<td>L201</td>
<td>2001</td>
<td>2009</td>
<td>9</td>
<td>2001</td>
<td>3.125 m</td>
<td>SW</td>
</tr>
<tr>
<td>L202</td>
<td>2001</td>
<td>2009</td>
<td>9</td>
<td>2579</td>
<td>3.125 m</td>
<td>SW</td>
</tr>
<tr>
<td>L301</td>
<td>3001</td>
<td>3009</td>
<td>9</td>
<td>3001</td>
<td>3.125 m</td>
<td>W</td>
</tr>
<tr>
<td>L302</td>
<td>3001</td>
<td>3009</td>
<td>9</td>
<td>3696</td>
<td>3.125 m</td>
<td>W</td>
</tr>
<tr>
<td>L401</td>
<td>4001</td>
<td>4009</td>
<td>9</td>
<td>4001</td>
<td>3.125 m</td>
<td>N</td>
</tr>
<tr>
<td>L402</td>
<td>4001</td>
<td>4009</td>
<td>9</td>
<td>4093</td>
<td>3.125 m</td>
<td>N</td>
</tr>
<tr>
<td>L403*</td>
<td>4001</td>
<td>4009</td>
<td>9</td>
<td>4093</td>
<td>3.125 m</td>
<td>N</td>
</tr>
</tbody>
</table>

* Supergather 403 was acquired with a hammer source. All other supergathers were acquired with the weight drop seismic source.

5.2 Data Processing

Seismic processing was done using the GLOBE Claritas software v6.3.1.12576 produced by GNS Science. A general processing sequence was established in chapter 2, section 2.3 from analysis of pilot data. Figure 5.3 outlines the basic processing sequence, and each component is explained in greater detail in subsequent sections. Information for each processing step was derived from Sheriff (2002), Yilmaz (2008) or Ravens (2001), and these will not be referenced again for this section, whereas other references will be stated. The post-stack processing is explained further within this chapter. The bulk processing of the RIS seismic lines was undertaken within Claritas using the job control system (JCS) where a single file (.jcs) with shot information, such as the number of CDPs per line, line orientation, and line number, were used to control the processing on a line-by-line basis.
Figure 5.3 Processing sequence similar to that found in chapter 2, with four post-stack processing steps (green, sections 5.2.3 to 5.2.6) that were applied to the RIS data, and are discussed in this chapter.

Table 5.4 Processing flow modules used to process the data to the final image. Module descriptions from the GLOBE Claritas Module List.

<table>
<thead>
<tr>
<th>Function</th>
<th>Module</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Job Control</td>
<td>SEISJOB</td>
<td>Defines project name, line ID, JCS parameters etc.</td>
</tr>
<tr>
<td>Data Input</td>
<td>DISCREAD</td>
<td>Reads SEG-Y or Claritas disc files.</td>
</tr>
<tr>
<td>Time Shift</td>
<td>STATIC</td>
<td>Application of elevation, residual, bulk statics etc.</td>
</tr>
<tr>
<td>Trace Editing</td>
<td>TREMOVE</td>
<td>Totally removes a trace from the processing flow.</td>
</tr>
<tr>
<td></td>
<td>TREDIT</td>
<td>Whole trace edit.</td>
</tr>
<tr>
<td>Adding Geometry</td>
<td>ADDGEOM</td>
<td>Adds geometry database information into trace headers.</td>
</tr>
<tr>
<td>Data Input</td>
<td>DISCSORT</td>
<td>Reads SEG-Y file and outputs sorted traces (no scratch file).</td>
</tr>
<tr>
<td>Velocity Model</td>
<td>NMO</td>
<td>Forward and inverse NMO with optional stretch mute.</td>
</tr>
<tr>
<td>Stacking</td>
<td>STACK</td>
<td>Stack of CDP (or shot) gathers.</td>
</tr>
<tr>
<td>Gain Control</td>
<td>AGC</td>
<td>Automatic Gain Control.</td>
</tr>
<tr>
<td>F-k Filtering</td>
<td>FKMUTE</td>
<td>Generalised muting of FK spectra.</td>
</tr>
<tr>
<td>Migration</td>
<td>FDMIG</td>
<td>Finite-difference post-stack migration.</td>
</tr>
<tr>
<td>Bandpass Filtering</td>
<td>FDFILT</td>
<td>Frequency-domain time and spatially varying filter.</td>
</tr>
<tr>
<td>Coherency Filtering</td>
<td>FXDECON</td>
<td>FX-domain complex Wiener deconvolution.</td>
</tr>
<tr>
<td>Data Output</td>
<td>DISCWRITE</td>
<td>Writes seismic traces to Claritas SEG-Y disc file.</td>
</tr>
</tbody>
</table>
5.2.1 Velocity Analysis

5.2.1.1 Root-mean Squared Velocity

The root-mean squared velocity \( (v_{rms}) \) was calculated from the interval velocities of single layers, and was calculated for multiple flat layers where the offset was assumed to be small compared with the depth. It is the velocity that was used if a multi-layer scenario would be simplified into a single-layer scenario, to give the same two-way travel time and NMO as observed in the seismic section (Mussett & Khan, 2000). From this, a hyperbolic moveout equation was derived as a power series in which \( v_{rms} \) was the velocity:

\[
v_{rms} = \sqrt{\frac{v_1^2 t_1 + v_2^2 t_2 + \cdots}{t_1 + t_2 + \cdots}}
\]

Equation 5.1 Root mean squared velocity (Mussett & Khan, 2000)

where \( t_n \) was the one-way travel time spent in each layer, and \( v_n \) the velocity of each respective layer.

The ice base was observed in the seismic data at approximately 204 ms as a seismically opaque section. From other data acquired at the same localities, the ice shelf was approximated to be 370 m thick in this area (W. Rack, personal communication, 2015). The root-mean squared P-wave velocity to the base of the ice was calculated using this thickness and travel time (equation 5.2):

\[
\frac{t}{2} = \frac{d}{v}
\]

\[v = 3627 \text{ m/s}\]

Equation 5.2 One-way travel time for a single layer.

\[
v_{rms} = \sqrt{\frac{v_1^2 t_1 + v_2^2 t_2}{t_1 + t_2}}
\]

\[v_{rms} = 3627 \text{ m/s}\]

The root-mean squared P-wave velocity calculated to the seafloor (using equation 5.1), used 3627 m/s as the P-wave velocity of ice, 1500 m/s as the P-wave velocity of water,
102 ms as the one-way travel time to the ice base, and 395 ms as the one-way travel time to the seafloor as seen on the seismic shot records:

\[ v_{rms} = \frac{v_1^2 t_1 + v_2^2 t_2}{\sqrt{(t_1 + t_2)}} \]

\[ v_{rms} = 2239 \text{ m/s} \]

5.2.1.2 Stacking Velocity

The stacking velocity \((v_{stack})\) was the hyperbolic velocity for a layer that best corrects for normal moveout, and was used in Claritas for stacking data (Figure 5.4). Where there were small offsets, or horizontal layers, \(v_{stack}\) should approximately equal \(v_{rms}\). In all four RIS lines, all features were horizontal, or near horizontal, and the stacking velocity was roughly equal to the root-mean squared velocity.

![Figure 5.4](image)

**Figure 5.4** Velocity model using calculated root-mean squared velocities for the seafloor, and was used for stacking the RIS data.

5.2.2 Final Stack

The final stack was produced once the velocity analysis was complete. The velocity analysis process was iterative, and continued until the seismic image was optimised. There is a large interpretive aspect to seismic processing, as the image quality is user defined, and can vary from minimal processing to get a rough image, to processing intensive images, where high accuracy is of importance. From this point, post-stack
processing modules were applied. There are a variety of stacking methods, and a conventional stacking method and un-normalised cross-correlation stacking were trialled (which attempted to make stronger events stand out more clearly). However, the conventional stacking method produced better results than the un-normalised cross-correlation method (Figure 5.5). This may be because of the large amount of coherent noise in the records.
Two different stacking methods used on Line 04, normal stacking (a) and unnormalised cross-correlation stacking (b), which attempts to make stronger events stand out more clearly. However, the normal stacking method created a better stack than the unnormalised stacking method, and will be used.
5.2.3 F-k Filtering

Some coherent noise types, such as ground-roll or surface waves, were more easily separated in the frequency-wavenumber (f-k) domain than the time-distance (t-x) domain, and were easier to mute when the inverse transform was applied. No other seismic events had velocities, or apparent velocities, as low as these surface waves, and f-k filtering easily removed these (Hess Corporation, 2015). The f-k spectrum/transform was the result of a 2D Fourier transform of the t-x domain. The wavenumber is defined as the number of cycles per unit of distance, which is similar to the idea of frequency in the time domain, where the peaks are per unit of time, so the wavenumber of an event is determined by the number of peaks counted within a unit of distance (Yilmaz, 2008). When seismic data were analysed in the f-k domain, the plot displayed energy density for a time interval and was plotted on an f-k graph (Figure 5.6). The colour intensity was used to show the energy density on the plot, and was helpful to indicate areas of energy that could be muted. This could be used to study the direction and apparent velocity of the seismic waves (Sheriff, 2002). Thus, the f-k plot characterised the 2D amplitude spectrum of the seismic section in the t-x domain. The spatial dimension was controlled by the trace spacing and needed to be sampled according to the Nyquist theorem (similar to sampling a time series).
Figure 5.6 Plots (a) that illustrate the conversion of three events (direct arrival, hyperbolic reflection, and horizon with no dip) from the time-offset (t-x) domain, to the f-k domain. On the right the events have been colour coded in both domains to better display how each event is seen in the f-k plot. The blue line is cut off on the right at the Nyquist wavenumber, and continues on the left-hand side (reproduced from Hess Corporation, 2015). The lower plot (b) is the f-k spectrum from the RIS data, Line 04. The f-k spectrum of data that is retained, inside black triangle, and the strong surface wave signal, highlighted in white, are shown on this image.

A horizon with no dip will have a constant wavenumber of zero. Horizons with the same dip, and constant apparent velocity, in the time-space domain, regardless of their location on the seismic profile, are mapped as a radial single line on the f-k plot.
(Yilmaz, 2008). Due to aliasing, this line left the plot on the right-hand side at the Nyquist wavenumber, and continued on the left-hand side. A steeper dip will result in the horizon aliasing at a lower frequency (Yilmaz, 2008). Surface waves on the ice, and seismic waves that purely travelled through the ice, were problematic in ice seismic data, and appeared on an f-k plot with a large wavenumber and a low frequency. In Claritas the FKMUTE module utilised an f-k filter, which in this case was created from the rawstack. The ice component of the RIS data displayed strong surface waves that were clearly seen in the f-k spectrum and could be removed, along with other noise. Figure 5.7 illustrates the effect of an f-k filter when applied to Line 04 and the extent of ice data and noise that is removed.
Figure 5.7 Rawstack of Line 04 (a) and the rawstack with an f-k filter applied (b). Note the large noise reduction in the second image, and the improved imaging of the seafloor. Events are more coherent and show up more clearly in the filtered section.
5.2.4 Migration

Migration is an operation in seismic processing which attempts to return dipping reflections to their original and true locations within the subsurface, and to collapse any diffractions. Migration was necessary as variable velocities and dipping horizons within the subsurface changed the location on the surface where they were recorded (Figure 5.8). An accurate velocity model was needed for migration, and the type of model varied depending on the migration method. The velocity model previously created (section 5.2.1) was smoothed to create a realistic velocity model that approximated a ‘true’ geological section. This new velocity model was then used as an input for migration processing. Sections with steeper dip angles or curved reflections required more migration. This process increased the spatial resolution of a seismic section and created a more accurate seismic image of the subsurface.

Figure 5.8 Zero-offset migration of a dipping reflector, where the red dots are the source/receivers, blue dots/line are the true geological dipping layer, and green dots/line the imaged version of (a) in a seismic section (which is plotted below at times equal to the receiver times, t1-t4). Note how the true dipping reflector (blue) is shorter and steeper than the imaged reflector. The dipping reflector on the seismic section is not the true reflector and interpretation of this would be incorrect. By taking a semi-circular arc equal to the travel time (to each of the recorded positions) and constructing a line at the tangent of the arcs, the line is migrated to its true position (Hess Corporation, 2015).
Post-stack, finite-difference migration was used on the RIS data, and was a method that efficiently implemented the downward continuation of migration in the time-depth (t-x) domain. It was a simple computer method used to calculate derivatives and second derivatives for the migrated section. Finite-difference time migration was able to handle some lateral velocity variation, and was able to handle data that had a low signal-to-noise ratio. However, it was relatively slow in speed to process, and had problems when dips were greater than 45°. In Claritas the FDMIG module was used for this migration, which used the previously created NMO velocity file (Figure 5.9). Migration also has the added benefit of reducing incoherent noise from the section by migrating or spreading the noise out in the stack, and was the aspect of migration that likely had most significant effect on the RIS data.
Figure 5.9 Line 04 pre-migration (a) and with the application of finite-difference migration, FDMIG, module (b). Both stacks have an f-k filter and AGC applied. Visually it is difficult to distinguish a large difference between the two sections; however, at closer inspection some of the horizontal events are more coherent in (a) then in (b), e.g. around CDP 700, at 800 ms.
5.2.5 Bandpass Filtering

See section 2.3.1.5. A bandpass filter with the corner frequencies 22, 45, 180 and 210 Hz was used on Line 04 post f-k filter and migration.

5.2.6 Coherency Filtering

The basic principle of coherency filtering is to compress the recorded wavelet and attenuate any reverberations to increase resolution and create a more accurate representation of subsurface reflections. Normally coherency filtering is applied before data stacking; however, it can also be applied after the data have been stacked. Random noise removal can be achieved through the use of frequency-space (f-x) deconvolution, which attenuates random noise using Wiener deconvolution in the f-x domain. The FXDECON module in Claritas is used for random noise removal, and was used on the RIS data (Figure 5.10).
Figure 5.10 Line 04 pre-coherency filtering (a) and with the application of frequency-space deconvolution, FXDECON (b). Both stacks have an f-k filter, AGC, migration, and bandpass filter applied. Visually it is difficult to distinguish a large difference between the two sections; however, at closer inspection some of the horizontal events are more coherent in (a) then in (b), e.g. around CDP 500, at 800 ms.
6. Results

This chapter presents the main seismic features observed and their geological interpretations. Four main reflectivity patterns (horizontal strata, pinch-outs, lap-outs, and deeper structures) were observed, from which two main seismic facies, an erosion surface, and deeper structures were inferred. The depths to/thicknesses of these seismic features were calculated; however, due to the resolution of data and variable velocity ranges, a margin of error is associated with these values.

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6.1 Observations

Depth conversions for the four seismic sections were done assuming horizontal layers, \( i \), using the following equation:

\[
h_i = v_i \left( \frac{t_{iB} - t_{iT}}{2} \right)
\]

**Equation 6.1** Layer thickness (Mussett & Khan, 2000)

where \( t_{iB} \) and \( t_{iT} \) are the two-way travel times to the base and top reflectors of layer \( i \), respectively, \( h \) the thickness of the layer, and \( v \) is the velocity of the layer. The travel time was divided by two as the time observed on the seismic line was the total time that it took for seismic waves to travel down to the reflector, and back to the receivers. The interval velocity model was used to calculate the depths to the sedimentary features, using 3627 m/s as the P-wave velocity of ice (chapter 5) and 1500 m/s as the P-wave velocity of water, and 1500 to 2000 m/s for saturated sediments, where 1800 m/s was the average value used for saturated sediments calculations (Mussett & Khan, 2000). Calculated depths are approximate only, as in reality, velocity variations were more complex than the simplified models assumed for these calculations. It should also be noted that each layer (i.e. the firn, glacial ice, water, and sediments) each have a range of associated P-wave velocities. This study has used either calculated velocities (no taking errors into account) or the average velocity (e.g. for the saturated sediments), where a large range of P-wave velocities is viable. Therefore, depth conversions should be analysed with some caution and be used as an approximate indicator of depth, not an absolute value.

6.1.1 Resolution of Observations

Four main features were observed within the first 200 ms below the seafloor, between 790 and 994 ms. An intra-ice multiple (i.e. where both the seafloor reflection and multiple path within the ice shelf are recorded) the was observed at approximately 994 ms. Below the intra-ice multiple other features were mostly obscured, which corresponded to a similar depth at which the ice-shelf base was observed at 204 ms. Possible deeper, dipping features are observed in lines 01, 02, and 03. The resolution of the RIS seismic data was limited by the Rayleigh resolution, which was
approximately equal to a quarter of the dominant wavelength, and corresponded to the smallest feature that can be imaged (Sheriff, 2002):

\[ \lambda = \frac{\nu}{f} \]

**Equation 6.2** Wavelength from velocity and frequency (Sheriff, 2002).

where \( \lambda \) was the dominant wavelength, \( f \) the dominant frequency, and \( \nu \) the P-wave velocity. The dominant frequency of the RIS data was 70 Hz, and a P-wave velocity of 2239 m/s (Figure 6.1; section 5.2.1) was used. This equalled a dominant wavelength of 32 m, and therefore, the thinnest layer that could be resolved in these data was approximately 8 m.

![Figure 6.1 Amplitude spectra graph of a raw shot in Line, which illustrates the dominant frequency of approximately 70 Hz.](image)

### 6.1.2 Seismic Features

Sequence stratigraphy analysis techniques were used to identify three main seismic features, as well as the seafloor, and consisted of pinch-out structures, lap-out structures, and deeper structures (Figure 6.2). It is highly likely that the data was collected in a sedimentary setting as opposed to a bedrock setting, as the seafloor was mostly horizontal, reflectivity was sub-horizontal, and coherent reflections were observed to extend approximately 200 m below the seafloor (64 times the shot spacing, see section 3.1.1 regarding depth penetration expected).
6.1.2.1 Seafloor and Horizontal Strata

The seafloor was observed as a strong reflection in all four seismic lines. It consists of two high amplitude cycles (i.e. the reflection was darker than other reflections), underneath which the seismic reflections were weaker, sub-horizontal, and undulated slightly. Figure 6.3 displays the characteristics of the seafloor.

![Figure 6.2](image1.png) Classic sequence stratigraphic seismic features which are used to identify different sedimentary structures in seismic lines (a). Both lap-out and pinch-out structures are present.

![Figure 6.3](image2.png) CDPS 700 to 1330, Line 03, illustrate the typical seafloor and horizontal strata expression observed between 790 and 825 ms, and characteristically consists of two high-amplitude cycles (i.e. the reflection is observed darker on the image than other reflections), underneath which the seismic signal is weaker. Red annotations indicate direction of lap-out or pinch-out features.
6.1.2.2 Pinch-out Structures

The pinch-out structures were defined by laterally discontinuous strata, with no clear lap-out structures present. These pinch-outs are observed multiple times within each seismic line (Figure 6.4).

![Figure 6.4](CDPs 100 to 280, Line 04, illustrate the typical pinch-out structures between 825 and 994 ms, which are defined by strata discontinuing laterally. Red annotations indicate direction of lap-out or pinch-out features.)

6.1.2.3 Lap-out Structures

The lap-out structures were defined by the strata appearing to either top-lap, on-lap, or down-lap a unit, differing in doing so from pinch-outs. Figure 6.5 shows the typical appearance of the lap-out structures.
Figure 6.5 CDPs 2670 to 3380, Line 03, illustrate the typical lap-out structures between 825 and 994 ms, which are defined by strata appearing to either top-lap, on-lap, or down-lap a unit. Red annotations indicate direction of lap-out or pinch-out features.

6.1.2.4 Deeper Structures

The deeper structures were only observed below the intra-ice multiple. They were defined by faint, dipping reflections. Figure 6.6 shows possible, faint reflections of the deeper structures observed in lines 01, 02, and 03.
Figure 6.6 CDPs 1250 to 1550, Line 03, illustrate the typical weak, dipping reflection of the deeper structures between 1255 and 1398 ms. Red annotations indicate direction of lap-out or pinch-out features.

6.1.3 Line 01

Line 01 was the lowest quality seismic line, because different shot methods were still being established early in the survey, and data at the centre of the line were of poor quality. At the start and end of this line, data were recorded on the second drop of the seismic source, whereas data in the centre of the line were recorded on the first weight drop of the seismic source. The first drop compacted the firn, which then reduced sound attenuation for the second weight drop. From this point forward all data were recorded at the second weight drop. Electrical noise is observed within this line, mostly at the beginning, and is distinguished by its high amplitudes, and upward curve (CDP 610, approximately 900 ms depth). The seafloor was sub-horizontal with slight undulations, and displayed horizontal strata within the first 35 ms below the seafloor. Strata could be observed to pinch out (CDPs 210 to 390, 2290 to 2530, 3210 to 3340, and 3730 to 3810, all between 825 and 994 ms), lap out (CDPs 1410 to 1510, and 2650 to 2790, observed between 825 and 994 ms), and there was evidence of deeper structures (CDPs 2330 to 2510, and 3570 to 3700, between 1300 and 1700 ms, and 1150 and 1200 ms, respectively). Areas of higher amplitudes were observed (e.g.
near CDP 2430, at approximately 950 ms). See table 6.1 for calculated depths and Figure 6.7 for the full seismic line.

**Table 6.1 Calculated depths for features observed in Line 01.**

<table>
<thead>
<tr>
<th>Feature</th>
<th>CDP</th>
<th>Two-way Travel Time</th>
<th>Depth</th>
</tr>
</thead>
<tbody>
<tr>
<td>Seafloor</td>
<td>-</td>
<td>790 ms</td>
<td>809.5 m</td>
</tr>
<tr>
<td>Horizontal Strata</td>
<td>-</td>
<td>790 - 825 ms</td>
<td>809.5 - 841 m</td>
</tr>
<tr>
<td>Pinch-out</td>
<td>210 - 390</td>
<td>825 - 994 ms</td>
<td>841 - 993.1 m</td>
</tr>
<tr>
<td></td>
<td>2290 - 2530</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>3210 - 3340</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>3730 - 3810</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lap-out</td>
<td>1410 - 1510</td>
<td>825 - 994 ms</td>
<td>841 - 993.1 m</td>
</tr>
<tr>
<td></td>
<td>2650 - 2790</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Deeper Structure</td>
<td>2330 - 2510</td>
<td>1347 - 1583 ms</td>
<td>1346.1 - 1582.1 m</td>
</tr>
<tr>
<td></td>
<td>3570 - 3700</td>
<td>1177 - 1239 ms</td>
<td>1176.1 - 1238.1 m</td>
</tr>
</tbody>
</table>
Figure 6.7 Line 01, un-interpreted final stack (a) and interpreted final stack (b). The interpreted line shows evidence of two possible separate seismic facies, separated by an erosion surface (red), two areas of bedrock structures, electrical noise (green), and the approximate location of the ice-water interface (blue). The total length of Line 01 is 11.6 km, and a larger version is found in Appendix I.
6.1.4 Line 02

Line 02 was of high data quality throughout, with none of the poor quality signal areas observed in seismic Line 01. As in Line 01, the seafloor was sub-horizontal with slight undulations, two cycles of sub-horizontal, and high-amplitude strata until 825 ms. There were pinch out features present (CDPs 1120 to 1250, 1530 to 1970, and 3770 to 3850, and 4140 to 4330, between 825 to 994 ms), and lap out features (CDPs 130 to 170, 550 to 670, 2060 to 2290, and 2710 to 3020, between 825 to 994 ms). There were multiple deeper structures visible below the intra-ice multiple (CDPs 550 to 670, 1060 to 1180, 3240 to 3310, and 3470 to 3580, between 1306 and 1373 ms, and CDPs 1620 to 1810, between 1249 and 1373 ms). See table 6.2 for calculated depths and Figure 6.8 for the full seismic line.

### Table 6.2 Calculated depths for features observed in Line 02.

<table>
<thead>
<tr>
<th>Feature</th>
<th>CDP</th>
<th>Two-way Travel Time</th>
<th>Depth</th>
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<tr>
<td>Seafloor</td>
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<td>809.5 m</td>
</tr>
<tr>
<td>Horizontal Strata</td>
<td>-</td>
<td>790 - 825 ms</td>
<td>809.5 - 841 m</td>
</tr>
<tr>
<td>Pinch-out</td>
<td>1120 - 1250</td>
<td>825 - 994 ms</td>
<td>841 - 993.1 m</td>
</tr>
<tr>
<td></td>
<td>1530 - 1970</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>3770 - 3850</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>4140 - 4330</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lap-out</td>
<td>130 - 170</td>
<td>825 - 994 ms</td>
<td>841 - 993.1 m</td>
</tr>
<tr>
<td></td>
<td>550 - 670</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>2060 - 2290</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>2710 - 3020</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Deeper Structure</td>
<td>550 - 670</td>
<td>1306 - 1373 ms</td>
<td>1305.1 - 1372.1 m</td>
</tr>
<tr>
<td></td>
<td>1060 - 1180</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>1620 - 1810</td>
<td>1249 - 1373 ms</td>
<td>1248.1 - 1372.1 m</td>
</tr>
<tr>
<td></td>
<td>3240 - 3310</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>3470 - 3580</td>
<td>1306 - 1373 ms</td>
<td>1305.1 - 1372.1 m</td>
</tr>
</tbody>
</table>
Figure 6.8 Line 02, un-interpreted final stack (a) and interpreted final stack (b). The interpreted line shows evidence of two possible separate seismic facies, separated by an erosion surface (red), four areas of bedrock structures, and the approximate location of the ice-water interface (blue). The total length of Line 02 is 14.5 km, and a larger version is found in Appendix I.
6.1.5 Line 03

Line 03 was the longest line, and the seafloor reflection was observed at approximately 790 ms along its length, and showed undulation, and sub-horizontal, high-amplitude strata for two cycles until 825 ms. The seafloor reflection was weaker at the end of the line. Pinch-out features were observed at CDPs 150 to 340, 710 to 900, 3820 to 4000, and 5430 to 5580, all between 825 to 994 ms. Lap-out features were observed at CDPs 2110 to 2470, 4620 to 4780, and 4890 to 5020, between 825 to 994 ms. Deeper structures were observed in five areas (CDPs 1290 to 1500, 2190 to 2340, 3560 to 3660, 4980 to 5060, and 5360 to 5530, between 1255 and 1398 ms). See table 6.3 for calculated depths and Figure 6.9 for the full seismic line.

<table>
<thead>
<tr>
<th>Feature</th>
<th>CDP</th>
<th>Two-way Travel Time</th>
<th>Depth</th>
</tr>
</thead>
<tbody>
<tr>
<td>Seafloor</td>
<td>-</td>
<td>790 ms</td>
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</tr>
<tr>
<td>Horizontal Strata</td>
<td>-</td>
<td>790 - 825 ms</td>
<td>809.5 - 841 m</td>
</tr>
<tr>
<td>Pinch-out</td>
<td>150 - 340</td>
<td>825 - 994 ms</td>
<td>841 - 993.1 m</td>
</tr>
<tr>
<td></td>
<td>710 - 900</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>3820 - 4000</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>5430 - 5580</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lap-out</td>
<td>2110 - 2470</td>
<td>825 - 994 ms</td>
<td>841 - 993.1 m</td>
</tr>
<tr>
<td></td>
<td>2670 - 3380</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>4620 - 4780</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>4890 - 5020</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Deeper Structure</td>
<td>1290 - 1500</td>
<td>1255 - 1398 ms</td>
<td>1254.1 - 1397.1 m</td>
</tr>
<tr>
<td></td>
<td>2190 - 2340</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>3560 - 3660</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>4980 - 5060</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>5360 - 5530</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 6.9 Line 03, un-interpreted final stack (a) and interpreted final stack (b). The interpreted line shows evidence of two possible separate seismic facies, separated by an erosion surface (red), five areas of bedrock structures, electrical noise (green), and the approximate location of the ice-water interface (blue). The total length of Line 03 is 17.4 km, and a larger version is found in Appendix I.
6.1.6 Line 04

Line 04 was the shortest line, and completed the irregular quadrilateral. It was of similar quality as lines 02 and 03, with no areas of poor data quality. The seafloor was sub-horizontal, high amplitude, and two layers of horizontal strata were observed between 790 and 825 ms. A pinch-out feature was observed at the beginning of the line, from CDPs 100 to 280, between 825 and 994 ms. No deeper structures were observed in this line. See table 6.4 for calculated depths and Figure 6.10 for the full seismic line.

Table 6.4 Calculated depths for features observed in Line 04.

<table>
<thead>
<tr>
<th>Feature</th>
<th>CDP</th>
<th>Two-way Travel Time</th>
<th>Depth</th>
</tr>
</thead>
<tbody>
<tr>
<td>Seafloor</td>
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<td>809.5 m</td>
</tr>
<tr>
<td>Horizontal Strata</td>
<td>-</td>
<td>790 - 825 ms</td>
<td>809.5 - 841 m</td>
</tr>
<tr>
<td>Pinch-out</td>
<td>100 - 280</td>
<td>825 - 994 ms</td>
<td>841 - 993.1 m</td>
</tr>
</tbody>
</table>
Figure 6.10 Line 04, un-interpreted final stack (a) and interpreted final stack (b). The interpreted line shows evidence of two possible separate seismic facies, separated by an erosion surface (red), and the approximate location of the ice-water interface (blue). The total length of Line 04 is 2.3 km, and a larger version is found in Appendix I.
6.2 Seismic Interpretations

Initial seismic interpretations were conducted on the seismic lines in regard to possible sedimentary facies (also see section 6.1.2); however, more work is still needed to identify source signature (i.e. the horizontal layers in seismic facies one could be an artefact of the source), and more work is needed on penetration of signal into the subsurface. An alternative interpretation for the lower seismic signal in seismic facies two was that the strata were caused by the signal in the ice shelf.

6.2.1 Sediments

The sediments beneath the seafloor appear to be broadly divisible into two separate seismic facies. Seismic facies one consists of the upper, horizontal layers of strata, and seismic facies two of the lower section of sediments, where strata are of a more irregular.

6.2.1.1 Seismic Facies One

Seismic facies one consisted of the upper two cycles of sub-horizontal strata, in approximately the upper 30 m below the seafloor, and its lower boundary was a possible erosion surface. These two layers are higher in amplitude on the seismic section, and likely represent sediments of a different nature to seismic facies two. This seismic facies possibly represents a mixture of sand and clay that have been deposited from the floating ice shelf, as opposed to a grounded ice sheet (Hambrey et al. 2002).

6.2.1.2 Seismic Facies Two

The second seismic facies contains the pinch-out, lap-out features, and is irregular in character than seismic facies one. This facies extends another approximate 150 m below seismic facies one and the erosional surface, at which point the intra-ice multiple obscures other sedimentary features. This seismic facies possibly contains till material, sands and gravels that were associated the grounded ice sheet, and with grounding zone retreat (Hambrey et al. 2002).
6.2.2 Bedrock
Lines 01, 02 and 03 all contain weak reflections that indicate deeper structures, all at approximately similar depths. Deeper dipping structures likely represent the top of the bedrock beneath the sediments, but are only faintly observed within the seismic data.

6.2.3 Erosion Surface
Between seismic facies one and two, there is a possible erosional surface, where the sedimentary characteristics change from horizontal layering, towards more complex and chaotic structures.
7. Discussion

This chapter addresses the primary objectives of this study, which were to characterise the seafloor bathymetry and its substrates, and constraining the nature of the ice shelf’s basal surface. The seafloor bathymetry was compared to both the RIGGS and ROSETTA data, and three possible hypotheses were suggested to account for observed seismic facies sediments and an erosion surface. A seafloor coring and sampling location is proposed from seismic data. The lack of distinct basal ice shelf interface reflection in the data were considered, along with other ice seismic processing issues encountered. A brief comparison of the synthetic shot records to with real world seismic data collected for this study is also made.

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7.1 Characterisation of Seafloor Bathymetry and Substrates, and Subsurface Depiction

The primary objective of this study was to characterise the seafloor bathymetry and substrates in the vicinity of the future hot water drill site, HWDS-2. Seismic data had to adequately depict the subsurface, to constrain decisions regarding future seafloor coring and sampling locations.

7.1.1 Seafloor Bathymetry

The seafloor is observed to consist of mostly horizontal reflections, with slight undulations, and reflectivity in the shallow sediments does not significantly vary laterally in depth. Interpolation from the RIGGS data proposes a broad (90 to 100 km wide) bathymetric trough and saddle structure within the vicinity of this study’s seismic lines (Bentley, 1984; Figure 7.1). The trough structure changes approximately 100 m in elevation over the scale of 50 km (from the centre of the trough to the edge), and at this scale, the dip angle of the seafloor would be approximately 0.1°. At the scale of this study’s seismic data, this angle is negligible and the trough and saddle structure is not observed at these dimensions (Figure 7.1). Nonetheless, it is a limitation of study that even though the seismic lines are multiple kilometres long (the longest being 17.4 km), this is a relatively short distance compared to regional and continental scale features. Thus, it is possible that these seismic lines have imaged part of the trough flank, but that this will not be visible. If this trough structure is present, then it is expected that sediments cored from the trough have a provenance from the WAIS as opposed to the EAIS, which is determined from ice flow mapping (Hulbe & Fahnestock, 2007).
The ROSETTA Project aims to characterise the ice, ocean, and bathymetry beneath the RIS. Initial ROSETTA gravity inversion data indicate a seafloor with greater relief than the RIGGS bathymetry (Tinto et al. 2014; Figure 7.2); however, the ROSETTA bathymetry indicates an elevation change of approximately 100 m over 20 km in the vicinity of the HWDS-2 seismic lines, which is a seafloor slope of approximately 0.3°. This is an angle too shallow to be visible on seismic data collected in this study. There are also two main limitations to comparing the ROSETTA data to the data presented here. The scale of the seismic data is much smaller than the ROSETTA data. As a result, this survey is not able to image long wavelength, regional changes. The second limitation is related to the nature of the two data sets. The gravity inversion data measure changes

Figure 7.1 Map of the interpolated trough and saddle bathymetry underneath the Ross Ice Shelf (RIS), with the seismic lines collected for this study (location indicated in red). The pink line indicated the current grounding line of the RIS (data points from the RIGGS survey, map produced by and used with the permission of C. Hulbe, 2016).
in the gravitational acceleration caused by variations in lithology densities to produce a gravity anomaly map (Mussett & Khan, 2000). Whilst this can be interpreted as changing seafloor bathymetry, the gravity anomalies may also indicate changes in lithology with a relatively smooth seafloor.

7.1.2 Substrate and Subsurface Geology

Both the substrate and the upper ~180 m below the seafloor likely consists of glacimarine sediments. The sedimentary strata are inferred to consist of two main seismic facies, upper horizontal strata, and a lower facies with irregular sedimentary structures. These two facies are separated by a possible erosion surface.

7.1.2.1 Sediments

Three scenarios are discussed that may explain the presence of two seismic facies. The first hypothesis infers that the thick sedimentary package to the boundary between facies one and facies two (approximately 30 m of sediments), represents the retreat of the Ross Ice Sheet during the LGM. The initial advance of the ice sheet would have eroded the uppermost sediments, and deposited grounding line fan sediments (Figure 7.3). Subsequent retreat of the ice sheet allowed for a period of sub-ice shelf or ice shelf sedimentary deposits, which was followed by another period of ice sheet re-advancement that eroded underlying sediments. It is a possible that facies one, facies
two, and the interface between represent a singular glacial cycle of deposition and erosion.

Figure 7.3 Typical glacimarine facies succession (Hambrey et al. 2002), with this example from the Ross Sea. The initial advance of the ice sheet erodes the uppermost sediments, and deposits grounding line fan sediments (a, b). Subsequent retreat of the ice sheet allows for a period of open water or ice shelf sedimentary deposits (b), which is then followed by another period of ice sheet re-advancement that erodes underlying sediments (c).

However, sedimentation rates in a glacimarine environment can vary, with McKay et al. (2008) suggesting a range of 1 cm/kyr to over 5 cm/kyr. Seismic facies one is approximately 30 m thick, and is hypothesised to have been deposited beneath a floating ice shelf since approximately 9 ka (McKay et al. 2016; Figure 7.4). This short time interval causes a discrepancy in sediment thickness if this glacial surface of erosion (GSE) is an indicator of the LGM (i.e. since the LGM at the above mentioned sedimentation rates, between 0.09 and 0.45 m of sediments would have accumulated, a large difference from the 30 m calculated). Therefore, it is unlikely that this GSE represents the LGM.
A second hypothesis suggests the seismic facies represent a longer period and a switch in glacial regime, from warmer, wet-base glaciation sediments underlying shallower and colder, dry-base glaciations sediments, during the Pliocene, where colder glaciation is the current state of Antarctica (Robesco et al. 2006). A regional erosion surface is observed in both the eastern and western Ross Sea (on the outer shelf), which marks a change in seismic character, and is estimated to have occurred around 3 Ma (Robesco et al. 2006). If the erosion surface extends beneath the RIS, between facies one and two, then the resulting sedimentation rate at the HWDS-2 locality would be approximately 1 cm/kyr, a similar rate to that observed by McKay et al. (2008) in sections of the AND-1B sedimentary core. Chronostratigraphic results from the AND-1B sedimentary core indicated a cold and relatively stable polar ice sheet in the late Miocene (between 13 and 10 Ma), which became more dynamic with significant subglacial stream discharges in the late Miocene-Pliocene (5 to 2.5 Ma), followed by periods of retreat, when the area experienced warmer oceanic conditions (between approximately 4 and 1.1 Ma). From this time to the recent, the ice sheet was again characterised by more stable cold polar conditions (Naish et al. 2007b). Though
there is a time discrepancy between Naish et al. (2007b) and Robesco et al. (2006), both studies support the shift from a warmer, wet-base ice sheet, to a colder, dry-base ice sheet; however, the timing by Naish et al. (2007b) is constrained by a sedimentary core, where as that of Robesco et al. (2006) is not. Nonetheless, it is likely that facies two may have been deposited during the Pliocene by warm, wet-based glaciers, and seismic facies one possibly from the mid-Pliocene to recent by cold, dry-based glaciers.

Lastly, it is a possibility that facies one and two are seismic artefacts, which result from the seismic signal within ice shelf, not the underlying geology. The observed seismic features, that are inferred to be sedimentary structures, could be ice shelf characteristics that were obscured by surface waves, but are visible as the intra-ice multiple. For example, forward modelling of synthetic shot records performed by Beaudoin et al. (1992) illustrates the proximity of an intra-ice multiple to the seafloor (Figure 7.5), which in appearance resembles facies one from this study’s data.

Further seismic analysis is needed, particularly on facies one, to determine whether the horizontal strata are a result of sedimentary features or seismic artefacts; however, even if the sub-seafloor reflections are artefacts, it is still very unlikely that these units are not sedimentary given horizontal nature of the seafloor.
7.1.2.2 Erosion Surface

From the above discussed sedimentary characteristics, the glacial regime change scenario is most likely and is carried forward in this discussion. Robertson and Bentley (1990) imaged a seismic reflector 50 to 150 m below the seafloor at RIGGS stations G9, I11, K11, and M9 (at depths of 110 m, 45 m, 156 m, and 54 m respectively), which has been interpreted as a glacial surface of erosion (GSE) within till, and was correlated to similar glacial erosional surface in the Ross Sea (Figure 7.6; Houtz & Meijer, 1970). The depth between seismic facies one and facies two in this study is calculated as approximately 30 m, and it is possible that the boundary between these two facies is a continuation of the same reflector observed in the RIGGS data.

![Figure 7.6](image-url) The seismic reflector was imaged at RIGGS stations G9, I11, K11, and M9 (light blue), and was present at 50 to 150 m below the seafloor. It has been interpreted as a possible glacial erosional surface within till material. The boundary between facies one and two within this study may be a continuation of the same reflector observed in the RIGGS data.

7.1.2.3 Bedrock
Low-amplitude, dipping reflections observed in lines 01, 02, and 03 may indicate bedrock structures, but they may also be seismic processing artefacts. With the WDSS, depth penetration is not expected to be sufficient to image features at this depth. If these features do indicate bedrock structures, they are at a depth (approximately 600 m below the seafloor) that will not affect future sediment coring in this locality. Dip direction is an indicator that these features are seismic artefacts, because the apparent dip of these features would change on a line-to-line basis due to the different line orientations.

A seismic refraction survey conducted by Robertson and Bentley (1990) at RIGGS stations BC, I10, J9, and RI (using explosive charges ranging from 91 to 458 kg) imaged sediments at a thickness range of 500 to 2000 m, which the sediments in this study are in range of (approximately 590 m). However, the WDSS it is not expected to image to depths of 2000 m because of the smaller amount of energy produced (Appendix II).

7.1.3 Seafloor Coring and Sampling Location Prognosis
All four seismic lines are similar in character (i.e. horizontal layering for the upper ~30 m, and irregular, dipping strata below an erosion surface); there is no individual area that is deemed a better target with respect to the seafloor and subsurface features. From this alone, it is recommended that the hot water drill site location should be established close to the South Pole Overland Traverse (SPOT) road and the 2015 season base camp, as this is a logical locality regarding logistics.

7.2 Characterisation of the Ice Shelf’s Basal Surface and Ice Features
The second main objective was to characterise the roughness of the ice shelf’s basal surface, which will in turn guide the deployment location of oceanographic equipment in future field seasons. This objective is not met directly from data in this study.

7.2.1 Basal Surface
The basal interface is not imaged as a distinct interface in this study. In the stacked seismic data, it is observed as a seismically opaque zone in the upper 200 to 210 ms, after which the signal changes character to low- to moderate-amplitudes in the water
column. The strength of the surface waves and ice signal in the seismic shots could be a factor that resulted in the obscuring of the ice-water reflection (Johnson & Smith, 1997); however, data from the ApRES indicated the presence of marine ice, and it is possible that a combination of the two factors has obscured this interface. The thickness from an autonomous phase-sensitive radio echo sounding (ApRES) measurement at the camp locality indicated an ice shelf thickness of approximately 370 m (W. Rack, 2015, personal communication). However, the lack of a distinct reflection and further ApRES data (M. Ryan, 2016, personal communication; Figure 7.7) indicated the presence of marine ice, which acted as a slushy zone and obscured the distinct reflection of an ice-water interface (Frankenstein et al. 2001; Mussett & Khan, 2000). Calculations from chapter 4 also indicate that the basal interface should only be observed under ideal conditions, with a strong ice-water interface and no accreted marine ice. Similar calculations can be performed on the RIS data where reflection and transmission values from the intra-ice multiple, seafloor, could indicate the amount of energy reflected and transmitted from the ice-water interface, and whether it is obscured by the surface wave data. However, this is outside the scope of this study, and should be conducted in future studies. From the data in this study, an assessment of the basal roughness cannot be made.
Figure 7.7 Autonomous phase-sensitive radio echo sounding (ApRES) measurement graphs from traverse (a) and campsite (b) localities. The graph from the traverse illustrates a distinct shift in amplitude at approximately 360 m, which is interpreted as a clear ice-water interface. The campsite locality illustrates a gradational change in amplitude, indicating the presence of marine ice in this area. (Graphs used with permission of M. Ryan, 2016).

7.2.2 Ice Seismic Processing

7.2.2.1 Intra-Ice Multiple

RIS data do display a relatively strong intra-ice multiple, which was modelled in the synthetic shot records (chapter 4). A study by Johnson and Smith (1997) investigating the seabed topography beneath the Ronne Ice Shelf displayed intra-ice multiples similar to those observed in this study’s data (Figure 7.8). It is unfortunate that the data do not contain a significant ice-water interface reflection, but multiples of this interface; however, this intra-ice multiple can be analysed to characterise ice properties, which require further processing that is outside the scope of this study.
Removing the intra-ice multiple from data was also problematic using standard filters (e.g. an f-k filter), as the moveout P-wave velocities of the multiple range from those of water to those of ice, a velocity range which also includes the expected velocity range of seafloor sediments (Beaudoin et al. 1992).

![Figure 7.8](image)

Figure 7.8 The intra-ice multiple seen within seismic data from the Ronne Ice Shelf (a; Johnson & Smith, 1997). Note the strong ice-water reflection seen in the Ronne Ice Shelf data, which is not seen in the Ross Ice Shelf (RIS) data.

### 7.2.2.2 Surface Waves

The data contain strong surface waves, which interfered with the seafloor signal. This strong ice signal was not observed in the MIS data that was processed to optimise receiver and source geometries. The MIS shot records were much cleaner (i.e. no ice interference) than the RIS data, where the ice component interferes with the seafloor reflection. The MIS lines were collected on ice that was approximately 82 m thick, compared to the 370 m ice thickness of the RIS data (Falconer et al. 2007). The thickness difference in the ice is the probable reason for the strong surface wave
presence in this study’s data in contrast to weaker surface waves in the MIS data. Both sets of data were collected at locations with similar seafloor depth (approximately 810 m on the RIS and 935 m on the MIS (Falconer et al. 2007)), and the varying ice conditions could be the reason for the larger surface wave interference in the RIS data. However, this could also be due to the different source types, where explosives were used as the source for the MIS data and the WDSS for the RIS data. A better analysis of the surface waves is required to characterise ice properties.

7.2.2.3 Flexural and Infragravity Waves
There was no evidence in this study’s data of either flexural waves or IG. It is possible that either type might be present; however, they have not affected the data in a way that has diminished data quality or the interpretations.

7.3 Data Acquisition System
The snow streamer and WDSS data acquisition system were a very effective method of data collection in the ice environment. This combined system allowed for rapid data collection, and the use of the snow streamer facilitated collection of 45.8 km of data, which could not have been achieved in the same time frame using spiked geophones. The acquisition system only required a small crew to operate, as opposed to a conventional land survey where larger crews are necessary to allow fast data collection. It is recommended that for future seismic studies of a similar nature, that the streamer and WDSS combination be used. Nonetheless, there are multiple issues specifically with the WDSS that require further considerations and improvement. The WDSS produces irregular double taps, where the weight bounced off the steel plate, and produced a secondary bounce aspect into the shot records. As this was an irregular secondary bounce, it is something that mostly stacks out in seismic data when the signal-to-noise ratio is increased. However, the use of deconvolution in processing could be another way of treating this problem, and is something that requires further investigation. Another downfall of the system was the problem of early triggering of the recording system. Approximately a third of all shots were triggered early by the same amount of time. Troubleshooting in the field did not diagnose the source of the early trigger, although some non-triggers were the result of soft snow which caused
the WDSS strike pad to float too high. Lastly, near the end of data collection, the wood and steel striking pad had reached the limit of its use, the wood had fractured in multiple places and the hardware connecting the wooden pad to the steel plate and trailer was becoming damaged. If longer seismic studies are undertaken with the WDSS, it is recommended that a secondary wooden striking pad is taken that can replace the first striking pad if it should fail.

7.4 Synthetic Shot Record Comparison

The synthetic shot records were useful in determining the limitations of data collection on the RIS, such as the 500 m or more ice thickness that would interfere with the seafloor signal (chapter 4). They were also useful in anticipating what shot records may look like, and what other features could be imaged. Both the synthetic shot records as well as the RIS shot records display strong ice components, a seafloor signal at approximately similar depths (790 ms in the RIS data, approximately 690 ms in the synthetic shot record for 350 m ice thickness), and an intra-ice multiple which obscures data below this multiple. Processing of the synthetic shots also allowed the advance development of seismic processing flows, which were then used in the field. A limitation of this process was the capability of the Claritas software to perform synthetic modelling. Its modelling capability is rudimentary, and cannot be performed without changing multiple parameters to values which are not realistic (e.g. the addition of more channels than the actual data acquisition system to avoid reverberations of seismic waves within the model window). For basic synthetic shot modelling, the Claritas module was adequate; however, for more extensive modelling software specifically dedicated to doing so is recommended.

7.5 Limitations

A limiting factor of this data was the vertical Rayleigh resolution, which is 8 m for the seismic data presented in this study. This vertical resolution limits the level of detail that can be seen in the sedimentary units, and needs to be taken into account when performing interpretations. All depths calculated will be accurate to ± 4 m, although this margin of error may increase because of errors in velocity assumptions. Another limiting factor of this seismic data was the strong signal of the ice shelf which
interferes with the seafloor and subsurface seismic signal. The ice shelf’s seismic response in itself warrants further study; however, for the purpose of this study, the ice shelf’s response imposed limits on the data quality. As described in chapter 5, multiple types of seismic filtering and processing steps were necessary to remove this signal and improve the image quality of the seafloor. Another factor that is limiting to this data is the inability to return to this locality to do more seismic studies. While collecting data in the field, it is possible to change the survey if features of interest are imaged; however, it is problematic, and most unlikely that a return to conduct more seismic studies would be possible, even if it is found that the processed data yields interesting features not seen in the field.
8. Conclusions

Our knowledge and resolution of the seafloor lying 323 km south of Scott Base and beneath the RIS has vastly improved since the days of the RIGGS data set, which until now provided the most detailed level of knowledge of this locality. Nonetheless, it is but a scratch on the surface of what could be known about this area. This study answered one of the two main research questions posed for this thesis, and has identified an appropriate seafloor coring and sampling location for future hot water drilling. The nature of the ice shelf base (whether it is rough or smooth) was not able to be determined from this data set. The use of the snow streamer facilitated collection of 45.8 km of data, which could not have been achieved in the same time frame using spiked geophones, and the synthetic shot records were useful in determining the limitations of data collection on the RIS.

The seismic data interpretation indicates over 180 m of sediments on the seafloor, which are separated into two seismic facies. Seismic facies one consists of horizontal, laterally continuous strata that are approximately 30 m thick. Seismic facies two underlies facies one, and consisted of irregular, dipping strata. The two facies are interpreted to be separated by an erosion surface. These two facies support a change in glaciation regime which is suggested to have occurred in the late Pliocene (approximately 3 Ma), where the lower sedimentary packages were deposited during a warmer, wet-based regime and overlying sediments were deposited by colder, dry-based glacial conditions. Future sedimentary coring in this area will be able to confirm or reject this hypothesis.

Multiple questions have arisen from this study, and these should be addressed in future research. The seismic data contain more information about internal ice shelf characteristics, and the data should be processed with this focus. The surface waves of the shot records will provide valuable information regarding ice characteristics, and should be treated accordingly. Future work on the basal roughness should include further analysis of the seismic data (e.g. detailed multiple analysis, surface wave
analysis, and reflection and transmission co-efficient value calculations) to determine the ice and ocean interface characteristics.

In closing, keep in mind that the main purpose of this seismic site survey was to characterise an area for future seafloor coring and sediment sampling. This is only the start of an interesting, multi-disciplinary project where this locality will be used for hot water drilling through the ice shelf, seafloor and sediment coring, deployment of an oceanographic observatory (instrument mooring), and collection of borehole ice samples and further seismic work focused on ice shelf characterisation.
9. References


Appendix I. Seismic Lines

Seismic lines 01, 02, 03, and 04 are available at the back of the thesis.
Appendix II. Weight Drop Seismic Source

The weight drop seismic source (WDSS) was purpose built for Antarctic seismic collection. The initial concept for the WDSS was developed from a similar seismic source designed by Marvin Speece at Montana Tech (Figure 1), which was used for land seismic surveys. Otago University utilised a hammer as their main seismic source for land seismic surveys. The weight drop seismic source aimed to reduce human error, increase the quality of the sound source, and gain deeper penetration into the subsurface than a hammer source.

The WDSS is a modified 250 kg fence post driver, mounted on a 9 by 4 ft caged trailer, and is powered by a hydraulic system and generator (Figure 2.). Purpose built skis are attachable for ice environment data collection.
Figure 2. The weight drop seismic source (WDSS) in Antarctica. Note the weight, hydraulic system, generator, and skis attached to the wheels.
An approximate calculation of the energy and impact force of the WDSS:

![Figure 3. Schematic of the weight drop, its potential energy (PE) at the top, and its kinetic energy (KE) when it hits the steel plate. Where m is the mass of the weight (kg), g acceleration due to gravity (9.81 m/s²), h the height by which the weight will drop (m), and v the impact velocity (m/s).](image)

\[ PE = mgh \]
\[ KE = 0 \]
\[ PE = KE \]
\[ mgh = \frac{1}{2}mv^2 \]
\[ KE = \frac{1}{2}mv^2 \]
\[ PE = 0 \]

where \( PE \) is the potential energy (Joules, J), \( KE \) the kinetic energy (J), \( m \) the mass of the weight (kg), \( g \) acceleration due to gravity (9.81 m/s²), \( h \) the height of the weight drop (m), and \( v \) the impact velocity (m/s). The above calculation makes multiple assumptions such as no air friction, and the friction of the steel weight against the steel I-beam was judged to have minimal effect on the energy and force calculations (lubricated steel on steel has a kinetic friction coefficient (\( \mu_k \)) of 0.05 (Knight, 2007) and has minimal effect over the short distance the weight is dropped.

The \( KE \) of 3,139.2 J gives us a rough value of the potential energy of the weight at the top converted to kinetic energy as the weight drops. To calculate the force (F) the work equation is needed:
\[ \text{Work} = Fs \cos \theta \]
\[ F = 31,392 \text{ N} \]

where \( s \) is the displacement (m) and \( \theta \) the angle between the force and displacement vectors. As the displacement is in the same direction as the force, the angle = 0. An approximate force calculation (Newton’s) gives a value of 31,392 N when a displacement of 0.1 m is assumed.

Compared to a hammer seismic source, where the hammer is generalised to be approximately 7 kg, hit from approximately 2 m high (depending on the person using the hammer), assuming gravitational acceleration, and a displacement of approximately 0.02 m, respective values of potential energy and force are calculated to be 137.2 J, and of 6,860 N.
Appendix III. Raw and Supplementary Data

A CD-ROM with raw and supplementary data is attached in the back of the thesis.