Tidal and Structural Controls on Seismic Events Near the Grounding Line at Beardmore Glacier, Antarctica

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TIDAL AND STRUCTURAL CONTROLS ON SEISMIC EVENTS
NEAR THE GROUNDING LINE AT BEARDMORE
GLACIER, ANTARCTICA

A Thesis
Presented to
The Graduate Faculty
Central Washington University

In Partial Fulfillment
of the Requirements for the Degree
Master of Science
Geology

by
Jade Linna Cooley

May 2017
We hereby approve the thesis of

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ABSTRACT

TIDAL AND STRUCTURAL CONTROLS ON SEISMIC EVENTS NEAR THE GROUNDING LINE AT BEARDMORE GLACIER, ANTARCTICA

by

Jade Linna Cooley

May 2015

Here I report seismic events occurring over a three-week period during the 2013-2014 austral summer near the grounding line of Beardmore Glacier, Antarctica. The ~24000 events over this time frame had a noticeable temporal pattern that correlates well with the principally diurnal tides of Antarctica. Falling and rising tide each accounted for nearly equal occurrence of events, and most (~42%) events occurred in the last third of any tidal cycle. Event epicenters were located using beamforming, and display a spatial pattern of two distinct clusters. Appearance of event location clusters differ on rising and falling tide. I theorize that, due to direction of glacier outlet and direction of overall ice shelf flow, the Beardmore Glacier can be separated into two zones near its grounding line that explain this pattern. There is an extensional zone which experiences most stress during falling tide, and a compressional zone that undergoes stress during both falling and rising tide.
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CHAPTER I

A. INTRODUCTION

The cryosphere's response to current changing environmental conditions is quick and extreme in comparison to other components of the earth system (Vaughan and others, 2013). Studying these responses allows for improved predictions of the consequences of climate change. The cryosphere is the part of the earth system containing water in its frozen state. This includes seasonal ice such as snowpack and river ice, as well as perennial ice such as glaciers and ice sheets. Ice sheets are large (>50,000 km$^2$), permanent expanses of ice that contain much of the ice and snow mass that can contribute to sea level rise. The planet currently has three ice sheets: Greenland in the north, and Antarctica, separated into two, in the south. The East and West Antarctic ice sheets are separated by the Transantarctic Mountains. The East Antarctic Ice Sheet (EAIS) is by far the larger, and is mostly grounded above sea level; the West Antarctic Ice Sheet (WAIS) is mostly grounded below sea level, which makes it far less stable (Weertman, 1974). The EAIS has the capacity to add ~60 m to sea level, but will take much longer to collapse due to its higher grounding level. The WAIS, inherently unstable due to being grounded below sea level, has the capacity to add 3.3 m to sea level (Bamber and others, 2009).

Both ice sheets are fringed by ice shelves, floating extensions of the ice sheets. Most of the ice sheets’ mass loss occurs in the form of basal melting and calving events, which take place on these ice shelves (Fig A.1; Joughin and others, 2012). The grounding line delineates the ice sheet, which rests upon bedrock, from the ice shelf. Grounding line retreat has been linked to several processes, and is a proxy for ice shelf thickness and
mass balance of the continent (Joughin and others, 2012; Rignot and Thomas, 2002). As ~74% of Antarctica’s ice mass exits the continent through its floating ice shelves, quantifying the mechanics at this transition leads to refined understanding of the ice sheet’s mass balance (Bindschandler and others, 2011). The amount of research investigating external driving forces of these mass loss mechanisms is limited, but further constraining these forces will prove useful in narrowing predictions of sea level rise for the coming years (Bassis and others, 2008; Barroul and others, 2013). These studies are not extensive, however, and additional observational data is required to improve

Figure A.1. Major Antarctic ice shelves.
Once ice is floating on the ocean it will displace its mass in the water, contributing to sea level rise. Thus, to get an accurate prediction of the rate at which Antarctica’s ice discharge will raise sea level, it is imperative to understand the speed at which its ice streams empty mass onto the ice shelves. Ice shelf stability is a major factor on ice stream speed, as these ice shelves provide an important buttressing effect on the ice streams by locally grounding on islands and sides of mountains along the ocean, restraining the grounded ice upstream (Fürst and others, 2016; Scambos and others, 2004, Joughin and others, 2012). This effect is greatly reduced when the floating ice shelf loses mass, such as through calving events, or becomes damaged, such as through crevasse formation. Responses are observed far upstream of the grounding line, where ice streams have been observed to speed up considerably, increasing the flux of mass exiting the continent (Scambos and others, 2004). The collapse of the Larsen B Ice shelf is a well-known example of this process (Scambos and others, 2004). The grounding line can be determined through a variety of ways, and in this study we use the MODIS grounding line, determined from manual inspection of ice shelf texture.

The Ross Ice Shelf is Antarctica’s largest ice shelf by surface area. Glaciers flowing from the EAIS through the Transantarctic Mountains, and ice streams from the WAIS’ Siple Coast both empty into this currently intact ice shelf. The Ross Ice Shelf has several larger glaciers along the Transantarctic Mountains, which empty catchments greater than 80 000 km$^2$ at rates greater than a meter a day (Stearns and others, 2011). These glaciers provide an opportunity to observe how glaciers attached to a large and
well-butressed ice shelf act. This study takes place on the Beardmore Glacier, that is one of these major glaciers. It has the benefit of being an area along the Transantarctic Mountains that is relatively well studied, and is where Shackleton and his group crossed the mountains in 1911 during the Terra Nova Expedition (Marsh and others, 2014).

A.1 Ocean Forces and Damage on Ice Shelves

The grounding line marks the boundary where the ice sheet moves from adhering to bedrock with high basal drag, to floating on the ocean with low basal drag. Here the edges of the moving ice shelf stick anywhere it is still in contact with the bedrock, causing deformation between the slow-moving areas near the grounding line, and the floating ice-shelf which is exposed to the movements of the ocean (Morlighem and others, 2010). Tidal forces subject ice in this area to shear, compressive, and tensile stresses, resulting in significant flexure (Smith, 1991). When the tide rises, the column of ocean water which presses against the ice shelf increases (Fig A.2). This enhances the ice shelf’s resistance to flow. The pressure applied by the ocean falls as the tide does, promoting increased velocity. In addition, rising tide will increase the volume of water underneath the ice shelf, lifting it slightly. At falling tide when this water retreats, the elevation of the ice shelf decreases. Eventually these stresses cause damage near the

Figure A.2. Simplified cartoon demonstrating the effect of rising and falling tide on ice shelves: compression during rising ride, extension during falling tide.
grounding line in the form of crevasses and rifts, often oriented perpendicular to the direction of flow. These are carried along with flow to eventually calve at the ice front (Bassis and Ma, 2015).

A.2 Cryoseismology

When crevasses form or icebergs calve off, they radiate elastic energy which can be recorded on seismometers. Recording and analyzing this seismic data gives insight into how these crevassing and calving events develop. The relationship between external driving forces and the location and timing of calving and crevassing events is poorly understood at present. Far from the grounding line ocean tides, ocean swell, and wind speed had little to no effect (Bassis and others, 2008). Near the grounding line, however, falling tide and thus extensional forces appear to have the biggest pull on crevasse formation (Barroul and others, 2013).

Icequakes associated with crevasse formation scale in magnitude with the size of the rupture, and so often these are relatively small and thus difficult to locate (Deichmann and others, 2000). In this study we use beamforming, useful for events lacking clear wave phase arrivals, to locate these small events. To further investigate the effect of falling and rising tide on these seismic events and thus crevasse formation, we look at the locations of events occurring during rising tide separately from locations of falling tide events. In addition to this, we look at the Beardmore Glacier in the context of the entire Ross Ice Shelf, and the forces which might act on it from merging with ice on the Ross Ice Shelf. We connect the external force of ocean tides with these structural constraints to explain spatial patterns of elastic energy release on Beardmore Glacier.
CHAPTER II

B. JOURNAL ARTICLE
TIDAL AND STRUCTURAL CONTROLS ON SEISMIC EVENTS
NEAR THE GROUNDING LINE AT BEARDMORE GLACIER, ANTARCTICA

B.1 Motivation

Antarctica’s ice sheets have the capacity to add over 60 m to sea level rise, and added ~0.41 mm/yr for the period of 2005-2010 (Vaughan and others, 2013). The two major processes by which Antarctica loses mass, basal melt and calving events, occur on the ice shelves which fringe the continent, so they are key locations for understanding its future contribution (Joughin and others, 2012; Fürst and others, 2016). These shelves also provide an important back stress on the ice streams which drain into them by locally grounding on islands and sides of mountains along the ocean, known as the buttressing effect (Dupont and Alley, 2005; Fürst and others, 2016). This buttressing effect is lessened by the formation of damage such as crevasses and rifts, and mass loss of the ice shelf such as calving. To be properly modeled and yield improved predictions of sea level rise, the forces which affect the ice shelves and damage them must be better understood. The grounding line which delineates the boundary of the grounded ice sheet and the floating ice shelf is an area of high deformation, and crevasses have been traced back to forming here (Bassis and Ma, 2015). As the grounding line retreats when the ice shelf thins and advances when it thickens, it can serve as a marker for ice shelf stability and thus predicting how much mass will be lost from the continent (Rignot and others, 2014).

Over three-quarters of Antarctica’s mass loss comes from its ice shelves, and the Ross Ice Shelf (RIS) is one of the largest ice shelves on the continent (Bindschadler and
others, 2011; Bentley and others, 1979). The RIS has a surface area of ~500 000 km$^2$, and serves as an outlet to the ocean for both the East Antarctic Ice Sheet (EAIS) and the West Antarctic Ice Sheet (WAIS). Ice is transported onto the RIS from the EAIS by glaciers along the Transantarctic Mountains, such as the Beardmore Glacier, and from the WAIS by ice streams along the Siple Coast, such as the Whillans Ice Stream. Major glaciers along the Transantarctic Mountains drain catchments of a variety of sizes, from 84 000 km$^2$ to 1 100 000 km$^2$, and the ice velocity over the grounding line ranges from 250 ma$^{-1}$ to 800 ma$^{-1}$ (Stearns, 2011). The Beardmore Glacier is a relatively well-studied major outlet glacier, which provides opportunities to look into ice shelf dynamics (Marsh and others, 2013; Marsh and others, 2014).

**B.1.1 Cryoseismology Implications**

Several glacial processes radiate elastic energy which can be recorded with standard seismic equipment. Analysis of the recorded elastic waves which take place on the ice or are generated by glacial processes, known as cryoseismology, allows for insight into the sources of this energy release. This technique has been proven to lend insight into mechanisms such as stick-slip motion (Winberry and others, 2011), brittle fracture (Bassis and others, 2007), and water-flow processes (Roeoesli and others, 2016).

Seismic observations have aided the study of fracture in ice shelves. GPS and seismic data acquired near rifts reveal propagation of these fractures might be observed as seismic swarms at regular intervals (Bassis and others, 2008). As rifts are linked to calving, this provides information on Antarctica’s mass balance (Heeszel and others, 2012). Research on a particular rift on the Amery Ice Shelf found wind speed, tidal
amplitudes, and ocean swell had no connection with the swarms of seismic events (Bassis and others, 2008).

Past research near the grounding line of Antarctica, however, has found a strong correlation between seismic events and tidal forces (Barroul, 2013). Analysis of seismic data taken at this location indicated temporal and spatial patterns of events, as well as epicenters which indicate areas where damage of the ice is likely to form. Relevant forces can be simplified into extension during falling tide and compression during rising tide. A larger mass of ocean water lifts and compresses the ice shelf, decreasing ice motion. Ocean waters retreating from the ice shelf during falling tide promotes increasing velocity along the length of the ice shelf from grounding line to ice front, as well as lowering elevation as the water in the basin below the shelf moves away (Anandakrishnan and Alley, 1997). Studying the different spatial patterns of falling tide events and rising tide events allows for insight into the different processes these tidal forces subject the ice shelf to, and their influence on ice shelf stability.

B.1.2 Study Area

Beardmore Glacier provides an excellent opportunity for cryoseismology research on grounding zone processes due its location and speed. It is one of the RIS’ major outlet glaciers, draining a catchment measuring ~90 000 km$^2$, flowing between coastal nunataks Mt. Hope (Fig B.1, grid S) and Mt. Kyffin (Fig B.1, grid N). The Beardmore moves at approximately 400 ma$^{-1}$ at the grounding line, between half and twice the pace of other major outlet glaciers on the RIS (Marsh and others, 2013). This medium pace allows for relatively safe in situ set ups of seismometer and GPS stations, while still being
representative of larger outlet glaciers. Previous research done at the Beardmore Glacier recorded detailed ice speeds, ice thickness (ranging from ~400-1150 m), vertical motion (0-0.55 m), and grounding line location measurements (Fretwell and others, 2013, Marsh and others, 2014). Ice velocity varies temporally both in daily and tidally punctuated cycles (Marsh and others, 2013). In this study we aim to present data on the spatial and temporal pattern of seismic events in conjunction with relevant forces to describe the effect of tidal forces on damage formation near the grounding line. This should aid further seismicity studies into ice shelf stability.

B.2 Methods

To record ice motion and seismic events, an array of GPS and seismic stations was deployed downstream of the grounding line on the Beardmore Glacier. Station array design and station spacing were determined largely through hazard-induced necessity, due to the heavily crevassed surface of Beardmore Glacier. At least one GPS and one seismic station were active from Dec 16, 2013 – Jan 7, 2014.

B.2.1 GPS Instrumentation

Five geodetic quality Trimble NetR9 receivers were set up 2-7 km downstream of the grounding line, to grid south of the center line of the glacier (Fig B.1). Three stations were spaced ~1 km apart, arrayed in a line perpendicular to flow. One additional GPS station was located ~5 km directly upstream of this line. Time series of station position were calculated using Precise Point Positioning methods (Zumberge and others, 1997) with the Natural Resource Canada’s web service. The derivative of the vertical motion of station G1 determines tidal velocity and thus falling and rising tide.
Figure B.1. Study area and station locations, black line is MOA grounding line a) Map of study area and all equipment locations. Seismic stations as blue triangles, GPS stations as red squares. Two GPS stations were located near enough to seismometers that their locations are combined as red triangles. Three stations nearest the grounding line were moved from three in the dense array, and active only for Dec 31-Jan 7. b) Close-up of relative station locations in subarrays; seismometers denoted by ‘S’; GPS stations denoted by ‘G’. Pink array is A1, purple array is A2, green array is A3. Centers of subarrays indicated by relevantly colored stars. Visible in grid N are crevasses that are the results of a submerged cliff of unknown material.

B.2.2 Seismic Instrumentation

Trillium 120 Q/QA broadband seismometers connected to Trimble REFTEK 130S-01 recorders were used for seismic data collection. A total of 13 stations were arranged in two geometries; the first for 14 days (Dec 16 – Dec 30, 2013) and the second for the remaining 8 days (Dec 31 – Jan 7, 2014). The first set up was a tightly spaced rectangle ~6 km downstream of the grounding line, and the second geometry involved moving three stations to an area ~2 km downstream of the grounding line (Fig B.1). Station spacing was approximately 1 km, which is roughly the thickness of the ice in this region (Fretwell and others, 2013). For later use in event location, three subarrays, A1-A3, were created. Subarray A1 consisted of seismometers S2, S3, S8; A2 consisted of
seismometers S9, S10, S12; and A3 consisted of seismometer S14, S15, S16. The centers of these arrays can be seen in Figure B.1.

**B.2.3 Event Detection**

To explore the temporal and spatial pattern of ice shelf damage, we monitored seismicity for 23 days. Analysis of raw seismic data revealed the data set was dominated by signals that emerged slowly from the noise, and so exact event initiation was difficult to distinguish. These events had no clear body wave arrivals and were of short duration (~3 s), with frequencies <20 Hz (Fig B.2). The amplitudes of events were highest in the vertical channel, and particle motion indicates predominately Rayleigh surface waves. As these Rayleigh surface waves indicate shallow processes, the two most likely event causes are surface crevasses and ice sliding across rock (Deichmann and others, 2000; Mikesell and others, 2012).

![Figure B.2. Spectrogram of 30 minutes of raw seismic data over all frequencies recorded by passive broadband seismometers. Most energy is below 20 Hz.](image)

The seismic time series was scanned for events using a short-term average/long-term (sta/lta) average moving window (Allen, 1978). This algorithm divides the amplitude average over a small section of the seismic wave by the amplitude average.
over a large section of time, with turn-on and turn-off thresholds defining a seismic event. Final parameters were determined from manual inspection of results for various values. In this study, we used a short time averaging window length of 2 s and a long time averaging window of length 120 s. Event start times were selected when the sta/lta value exceeded 2.7 and event termination was defined to be when the ratio fell below 1.7, to minimize catching multiple events occurring close together counting as one. Event waveform data and associated start times and end times were saved into a catalog for future use in locating events and correlating them to tide.

**B.2.4 Event Location**

Since the events on Beardmore Glacier are emergent, traditional seismic location methods that rely on seismic inversion from well-picked phase arrivals are problematic. While our seismic arrivals lack identifiable phases, the emergent signals often show similar waveforms across the subarrays. Correlating the time offsets of these similar waveforms, or beamforming on surface waves, can thus be used instead of traditional triangulation methods. Beamforming is common in solid-earth studies, that have used it to locate seismic swarms and small magnitude events (Inza and others, 2011; Pezzo and others, 1997). Beamforming gives likely velocity of event waveforms as well as azimuths, and is mostly sensitive to events outside of an array. In contrast to traditional location methods, the beamforming method will result in an azimuth, but not a distance; similar to traditional methods, multiple measurement points are required to triangulate the best-fit event location. Thus to enable event location, we separated our complete array of
13 stations into the subarrays A1, A2, and A3. The stations in these subarrays must be close enough that dispersion does not distort the similarity of the waveforms.

Beamforming uses waveforms recorded on multiple seismometers within an array. To determine the azimuth and velocity of an event, we iteratively determine the time shift required to align the waveforms. This is achieved by calculating the travel time to all stations from an assumed azimuth, with an assumed velocity. To get a single azimuthal measurement, the beamforming method goes through the following 3 steps for azimuths at 0-360° around the center of the stations being used.

1) Calculate the time lag between stations, assuming a specific slowness (reciprocal velocity) and direction the wavefront is traveling from. Shift the waveform at each station by this predicted time lag.

2) Compare envelope functions of shifted waveforms by measuring the correlation coefficient

3) Repeat for an appropriate range of slownesses.

Figure B.3 demonstrates the application of these steps for our three subarrays, over the same event.

1) For a given event, we assume a planar wavefront traveling across a two-dimensional seismic array from a specific direction. This wavefront travels at some slowness $S$ with Cartesian components $S_x$ and $S_y$. From this slowness, the relative arrival time $t_{ij}$ between the $i$th and $j$th station are calculated with:

$$t_{ij} = S_x x_{ij} + S_y y_{ij}$$  \hspace{1cm} (1)
Figure B.3. Beamforming method applied to the three subarrays used in this study. This is event #20032, on Jan 2\textsuperscript{nd} at 2:08 pm.  

a-c) Waveforms as they arrived at each station in subarray. Dispersion can be seen between the waveforms near the event (S2, S3), and those further from it (S14-S16).  

d-f) The overlap which corresponds to highest correlation coefficient determined from Eq (1).  

g-i) Probability maps built by searching through azimuths for all discrete values of a range of slownesses. Speed indicated by radius: blue circle at 1.6 km/s, green circle at 4 km/s.  

(Mori and others, 1994), where the $i$th stations is a distance from the $j$th station determined by Cartesian coordinates $x_{ij}$ and $y_{ij}$. The difference in the calculated arrival times between stations is the predicted time lag which would result from a wavefront moving at the slowness and from the direction used for the calculation. The waveform at station $i=1$ is not shifted, and the waveforms at all others stations in the subarray are then shifted by their calculated time lags relative to the reference station.
2) After shifting, waveforms may overlap to some degree. If the correlation coefficient of a specific overlap is near 1 (Fig B.3.d-f), it indicates the slowness and direction used to calculate the relevant time lags in (1) is likely the slowness and direction of the physical wavefront for that event. If the waveforms do not overlap, or the overlap is weak (Fig B.3.a-c), this results in a low correlation coefficient, and thus a low likelihood of the physical wavefront moving at that slowness from that direction.

Calculating Eq (1), shifting of waveforms, and recording the correlation coefficient of that shift is repeated for every value in a grid search, for all azimuths 0-360°.

3) Steps 1 and 2 are repeated for discrete values of an appropriate range of slownesses. Each value of slowness is used in Eq (1) and for all azimuths 0 – 360°. These results are displayed as probability contours with correlation coefficient of waveform overlap reflected in intensity of color, radius indicating slowness, and the angle indicating direction to event epicenter from the center of the array, with N at 90° (Fig B.3.g-i). To be more useful, slowness can be converted to horizontal velocity by:

\[ V_h = \frac{1}{\sqrt{S_x^2 + S_y^2}} \]  \hspace{1cm} (2)

With this, circles in the probability map are of radius equal to horizontal wavefront velocity as seen in Figure B.3.g-i.

Steps 1-3 of beamforming will result in a most likely azimuth and slowness, corresponding to the point that represents the highest correlation coefficient of shifted waveforms for a given subarray. The radius at this point indicates most likely slowness of the wavefront for that event. To determine the likely direction the wavefront came from,
back azimuth is calculated for this point from:

\[ \phi = \arctan\left(\frac{S_x}{S_y}\right) \]  

(3)

This indicates the direction from the center of the array of stations, to the most likely event epicenter.

The beamforming method yields a single value for back azimuth. To address uncertainty inherent in this measurement, we treat the azimuth of each subarray as a ray that broadens to take up 40°. The center of the azimuth, through the point determined from Eq (3), represents a weight of 1. This azimuthal ray weight strength falls in a linear gradient on either side of the center line to 0 at ±20° (Fig B.4.a-c). Thus, the azimuth determined by the beamforming method is the most likely direction from which the wavefront originates, but there is a chance > 0 that the event took place up to 10° on either side of that line.

Applying beamforming steps 1-3 to each subarray, for each event separately yields a likely azimuth for each subarray. Using the intersection of all individual rays from all subarrays provides an estimate of event location. Each azimuth is mapped as a line beginning at the center of its respective subarray. Multiple lines, and accordingly multiple rays, calculated from beamforming are mapped onto one image. These rays might intersect for a given event, and this intersection indicates the likely epicenter (Fig B.4.d).

In addition to ensuring spacing of stations in a subarray was minimal, we took relative location of the centers of these subarrays into account. If the centers of subarrays
Figure B.4. Intersection of rays for event #20032, a well-located event. Graded probability rays for azimuths of subarrays a) A1 (pink) b) A2 (blue) and c) A3 (green). d) Outlines of weighted azimuth rays from a-c, with intersection colored in same probability scale. Likely epicenter is calculated as the location of highest robustness, or sum of weights of the rays.
are spaced in a line – such as if a subarray consisted of S4, S5, and S7 in addition to A1 and A2 – beams may become nearly parallel with small differences between azimuth. This can result in a large intersection, which gives a large range of distances and thus location and magnitude. Thus, for the purposes of this research, we located events only when S14, S15, and S16 were active. Creating a subarray with these stations in conjunction with A1 and A2 creates centers of subarrays in a triangular pattern (Fig B.1.b). In addition, the distance from A3 to A1 and A2 is larger than the distance from A1 to A2. This results in more broadly spaced centers of subarrays to calculate back azimuth from.

A3 was active Dec 31st – Jan 7th, which allows for location of ~33% of events. Grid resolution for the 360° azimuth search in Step 2 of the beamforming method was 0.2 km, for a square of 25 by 30 km. Velocities of 1-6 km/s (slowness of 1/1 – 1/6 skm⁻¹) were combed in a resolution of 0.1 km/s⁻¹. To compare events between stations, we calculated a smoothed energy envelope function by taking the absolute value of amplitude and applying a 0.5 s running mean.

Occasionally event locations can be skewed by systematic errors such as an event being relatively noisy, or two events occurring very close in time relative to their duration. To ensure events such as these are not used for mapping locations or in magnitude calculations, we enforced the following conditions to define an event as “well-located”:

1) The rays of all three subarrays overlap.

2) The sum of azimuthal ray weights (robustness) must be greater than 2.1.
3) The intersection area is < 5 km².

All three subarray azimuthal rays should be overlapping, to ensure a higher
certainty in event location. Robustness is calculated by summing the value of the
azimuthal ray of all three subarrays at all points. Where there is no overlap, robustness is
0. In the case that only two of the three subarrays create an area of intersection, while the
third subarray is either not active or pointing in a direction which cannot overlap with the
other two, the maximum sum of azimuthal ray weights can be no greater than 2. Addition-
ally, all three rays can overlap on the edge of the linear gradient, creating an
intersection of very low robustness. Therefore, we use only event locations that are the
result of a robustness > 2.1.

A smaller area of intersection suggests smaller errors in location; thus condition
3) ensures a smaller error, as well as culling far-off events. If an event is far, noisy, or one
event is counted where two happen on the tail end of each other, azimuths from all
subarrays can end up being nearly parallel. After applying the ± 20° ray weighting, the
area of intersection of the three nearly parallel rays can be very large, sometimes
resulting in a parabola without ends that are bound within the grid of our search. Locating
only events with the smaller area of intersections avoids locations which might have a
large range of possible distances. This puts an upper limit on the error of our location: ±
2.5 km. This will also cull large (M_L>2) events. These events are recorded from far
distances (>15 km), that results in rays running nearly parallel.

These errors in location can be attributed to the emergent arrival nature of the
seismic events, the close spacing of the stations, and the difficulties inherent to picking
sta/lta parameters. Due to this ± 2.5 km error in event location, we focus on broad patterns of event epicenters in this study, and put less importance on exact epicenter locations.

To better quantify the area within which we will find well-located events from these conditions, we use a synthetic source algorithm. We conducted a grid search around subarrays, assuming a well-located event epicenter was in that grid point. Based on this, we can deduce how rays will overlap from all three subarrays. When setting the intensity to 2.1, the last condition is area of subarray ray intersection. When this area is 5 km$^2$, only locations within a few kilometers around the stations will contain well-located events (Fig B.5). Increasing this area to 25 km$^2$, which would inherently have a large error in distance, adds a small section to grid W and grid E that might also contain well-located events. As this area had very little seismic activity, we kept condition 3) at area of subarray ray intersection < 5 km$^2$. This method assumed perfect waveforms, which is why in our physical data some waves are located outside of the ranges shown available by this method.

**B.2.5 Event Magnitude**

Local magnitude estimates were calculated from:

$$M_L = \log_{10}(A_o)$$

(4)

where $A_o$ is the amplitude at the origin. $A_o$ is estimated from an equation which relates maximum amplitude of the waveform signal at all stations used in the subarrays and the amplitude at origin by a decaying function:
Figure B.5. Expected locations of well-located events. Considering a perfectly located events and a robustness > 2.1, a) area in which subarray ray intersection < 5 km$^2$. b) area in which subarray ray intersection was increased to 25 km$^2$.

\[ A_i(R_i) = A_o e^{-\alpha R_i / \sqrt{R_i}} \]  \hspace{1cm} (5)

Where $A_i$ is the maximum amplitude of the signal at the $i$th station, $R_i$ is distance from the $i$th station to event location, and $\alpha$ is the attenuation decay constant that describes how amplitudes decay with distance (Mikesell and others, 2012).

This equation is solved numerically. To solve for $\alpha$, an exponential function is fit to the graph of amplitudes and distance at each station for a given event (Fig B.6). This process is repeated individually for each well-located event. Amplitude should decay with distance, but occasionally errors occur, such as when an event takes place between and roughly equidistant to all subarrays. This results in amplitudes with very little difference between them, and fitting an exponential function to a straight line can result in a positive alpha, which doesn’t make physical sense. Based on the range of physically plausible values found in earlier studies, we use only events with $-0.2 > \alpha > -0.8$ to
Figure B.6. Determining alpha for a single event. This is the exponential fit for calculation of alpha for event #20032. Amplitude at each station is graphed vs that station’s distance from the event epicenter, and the exponential fit gives an alpha = -0.23.

calculate magnitude (Mikesell and others, 2012).

Source and station correction coefficients are set to 0, which means magnitudes are relative and not connected to an established scale. Using these relative measurements of magnitudes reveals patterns in the general size and range of sizes of magnitude of events. Magnitude errors will scale with location errors.
B.3 Results

B.3.1 GPS Derived Results

Horizontal velocity of each GPS stations was tidally modulated, similar to that seen in previous GPS velocity data near the grounding line of the Beardmore Glacier, with highest velocities on falling tide, and lower velocities on rising tide (Fig B.7; Marsh and others, 2013). Spring and neap tide are reflected in GPS velocities as well, with spring tide resulting in larger fluctuations of velocity and neap tide resulting in smaller fluctuations. The stations located ~6 km downstream of the grounding line (S1, S2, and S3) experienced changes on the order of ~0.1 m/day in horizontal velocity. Surprisingly, the station nearest the grounding line (G4) experienced larger fluctuations, on the order of ~0.7 m/day. This results in G4 having a velocity ~0.3 m/day faster than downstream stations during falling tide, and thus a shortening of distances between upstream and downstream stations. At rising tide, downstream stations move ~0.2 m/day faster than G4, and distance between G4 and downstream stations is lengthened.

We observe tidally modulated changes in the elevation of GPS stations as well. Similar to previous analysis of satellite imagery data at the Beardmore Glacier, we recognize that stations are not in equal degrees of hydrostatic equilibrium (Marsh and others, 2014). If these stations were in hydrostatic equilibrium, they would be in balance with the external force of the ocean, and changes in elevation of the ice and thus the stations would match the height of changes in elevation of the ocean water itself. In contrast, stations not in hydrostatic equilibrium have decreased changes in elevation, suggesting they are mechanically coupled to nearby mountains or grounded ice. Station
G3 experienced slightly less elevation change than station G1, which is located ~2 km further away from the Mt. Hope where the ice shelf is grounded. These elevation changes differ by ~3 cm at peak displacement (Fig B.8).

Figure B.7. Tidal variation in horizontal velocity of GPS stations. Change in speed for all stations correlate to changes in ocean tidal cycle. Stations move slower during rising tide (highlighted in red), and faster at falling tide (highlighted in blue). G4, nearest the grounding line, experiences the most extreme changes.

G3 experienced slightly less elevation change than station G1, which is located ~2 km further away from the Mt. Hope where the ice shelf is grounded. These elevation changes differ by ~3 cm at peak displacement (Fig B.8).
Figure B.8. Change in elevation between a station near (G3) and a station further (G1) from area of low flexure. Tidal offset of vertical change in motion implies one station moves less than another located only a few kilometers away. This being height of G3 subtracted from height of G1, we can see G1 has a more extreme tidal modulation and thus the ice nearer Mt. Hope experiences less vertical motion.

B.3.2 Event Detection

For our sta/lita parameters, we detected 24028 events over the 23 day (551 hour) period from Dec 16 – Jan 7, averaging 44 events per hour. These seismic events typically last between 1 and 10 seconds, with a median value of 3 seconds.

The rate of occurrence of seismic events was not constant: twice daily there was a steady increase to a noticeable peak in seismicity that lasted 2-3 hours. This was followed by an abrupt decrease to a rate 20-50% that of the peak seismicity (Fig B.9). When
compared to the predominantly diurnal tide of Antarctica, there was an obvious correlation. For any given tidal cycle the number of events steadily increases, with ~42% of events occurring in that tidal cycle taking place during the three hours prior to the tidal height extremum. The three-week data span included both spring and neap tides, and we observed that there was a higher number of events during the higher tidal range of spring tides, and lower peaks of number of events during the more subdued tidal height variations of neap tide. Seismic events that occurred during falling tide constituted 54% of total seismic events, while rising tide accounted for 46%.

![Graph showing temporal pattern of events by 3 hour windows. Last ~3 hours of any given ocean tidal cycle correlates to peak in seismic activity. Falling tide sees slightly higher peaks than rising tide. Part of a longer period (~1 month) tidal cycle of spring and neap tides is observed. This is over a selected portion of the time frame meant to illustrate the effect of spring and neap tide. Relative tidal height added for clarity, no attached scale but maximum peak to trough height is 70 cm.](image)

**B.3.3 Event Location and Magnitude**

Of the 8028 events which occurred in the last week of the data collection, one quarter (1783) are considered well-located. These event locations display a spatial pattern of two distinct clusters (Fig B.10). We will refer to these clusters according to their direction in polar stereographic terms, or the grid N and grid S cluster. The grid S cluster constitutes 66% of well-located events, covers an area of 26 km², and is located...
Figure B.10. Distribution of events, their magnitude and duration. A) Histogram of duration of events spanning full data set. B) Histogram of magnitude only from well-located events with alpha $-0.8 < \alpha < -0.2$. c) Spatial pattern of well-located events with good $\alpha$. Events have been dithered for visual clarity. Clusters labeled as (grid) N and (grid) S.
completely downstream of the grounding line. The grid N cluster constitutes 25% of well-located events. It is slightly more elliptical in shape and has a larger area of 49 km$^2$, it also spans the grounding line. The remaining ~9% of well-located events are located outside these clusters. As both these clusters lie near the center of the glacier, and are several kilometers from an ice-rock interface, the surficial seismic waves are more likely due to crevasse formation than ice sliding across rock.

Events that are both well-located and have -0.2 > $\alpha$ > -0.8 were used to calculate magnitude. This constitutes 1366 events, or ~77% of the well-located events. The magnitude of events is very small, and there is not a large range of magnitudes. Mean magnitude was $M_L = -1.7$, and most events occur within the range of -3 < $M_L$ < 0 (Fig B.10). There is no discernable pattern between magnitude and location.

**B.3.4 Event Location and GPS Displacement Spatial Patterns**

The 8028 events from Dec 31 – Jan 7 are representative of the complete data set of 24028 events in that the number of events in falling and rising tides are comparable. Similar to the temporal pattern, the spatial pattern of events is tidally controlled. Rising tide events are more likely to occur in the grid S cluster, and falling tide events are more likely to occur in the grid N cluster (Fig B.11.a-b). Over half (58%) of falling tide events are located in the more spread out N cluster, and an additional 30% are in the smaller S cluster. Rising tide events are more centrally located, with 89% occurring in the S cluster and only 4% in the 49 km$^2$ region of the N cluster.

Tidal controls on velocity are derived from displacement of GPS stations. Stations experience greater displacement during the extensional force of falling tide, which is
expected from falling tide as the extensional source. On rising tide these stations move much slower (Fig B.11.c). Stations nearer buttressed ice, such as the grounding line, are expected to experience smaller fluctuations in horizontal velocity due to the dampening effect of the grounded ice on motion. However, the station nearest the grounding line,

![Diagram](image)

**Figure B.11.** Rising and falling tide location patterns. a) Locations of events occurring during falling tide. b) Locations of events occurring during rising tide. For c) and d) histograms of well-located events in grey, Tidal height black dashed line, falling tide highlighted in blue and rising tide highlighted in red. Day 364-367 is the time period during which stations G1, G2, G4, and S14, S15, S16 were active. c) Green line is the change in location of station G1, white triangle in a), tidally modulated and moving slowly downstream. d) Green line is horizontal distance between stations G2 and G4, white triangle and circle respectively in b). Distance is tidally modulated, with stations moving further apart during rising tide.
G4, experiences the highest fluctuations. A result of G4’s larger fluctuations in velocity compared to other stations is G4 moving progressively closer to downstream stations. This is unexpected from the theory that ice nearer the grounding line should be more highly buttressed than ice further downstream. In addition, station G4 and stations downstream get closer together during falling tide (the extensional force), and further apart during rising tide (the compressional force).

**B.4 Discussion**

Some results from this study reinforce accepted ideas on extension and crevasse formation near the grounding line during falling tide. The quantity and tightly spaced location of rising tide events were unexpected. We use GPS velocities and relative positions to investigate the kinematics of the Beardmore Glacier, to explain the mechanism causing these rising tide events.

Our seismic network recorded ~24,000 events over three weeks. These low magnitude, emergent events had a temporal and spatial pattern. Temporally, seismicity steadily increases during any tidal cycle, with 42% of events in a single tide occurring during the three hours preceding tidal height extremum (Fig B.9). Similar to previous research done on the Mertz Glacier, we observe peaks of seismicity on falling tide (Barroul and others, 2013). Dissimilar to this previous research, we observe a nearly equal number of events during rising tide. The forces acting during rising tide and falling tide are comparable in intensity, with falling tide accounting for slightly more events over the entire period, but rising tide accounting for more well-located events during the Dec 31 – Jan 7. Spatially, event locations occur in two distinct clusters on the ice shelf.
Rising and falling tide create distinct patterns in locations of events on the Beardmore Glacier (Fig B.11).

The Beardmore Glacier lies between two mountains, with an asymmetric bed topography between them (Fretwell and others, 2013). This topography leads to the ice traveling preferentially closer to Mt. Kyffin, where the bed is deeper. This is reflected in the ice speed, with fastest ice velocity downstream of the grounding line to grid N, towards Mt. Kyffin (Fig B.12). This ice velocity and the ice thickness which results from it does not fully explain the pattern of seismic events. It does suggest we should investigate the forces that result from this pattern of ice velocity, and look at the Beardmore Glacier in the context of the larger Ross Ice Shelf.

B.4.1 Explanation of Clusters

The Beardmore Glacier discharges ice onto the Ross Ice Shelf, with velocity that is traveling in a grid SW direction. Ice on the Ross Ice Shelf upstream of this point, however, travels in a grid SE direction. The ice the Beardmore discharges then merges with ice moving from upstream on the ice shelf, and thus the ice from the Beardmore Glacier must change direction towards the ocean. This change in direction of travel means ice moving over the grounding line of the Beardmore Glacier curves to grid S only ~20 km downstream of the study area (Fig B.12). This creates a visible “elbow” in the ice mass, which can be seen in satellite imagery (Hulbe and others, 2007). Ice on the outside of the elbow undergoes extension in traveling around this bend, while compressing ice on the inside of the elbow into Mt. Hope. Compressional features can be seen inside the elbow, as buckles downstream of the S cluster. Ice on the outside of the
curve moves around this more buttressed ice, and thus undergoes extension. We will refer to these areas as the “extensional zone” – ice on the outside of the elbow, and the “compressional zone” – ice on the inside of the elbow (Fig B.13). The extensional zone further stresses on the compressional zone by pushing it towards Mt. Hope, adding to the dynamic force of ocean tides and creating three spatially and temporally distinct clusters:

![Figure B.12. Ice speed with well-located event epicenters as black dots. Dashed line through fastest ice speed. Curved change in path of fastest ice visible, taking a sharp turn grid S ~20 km downstream of event clusters. Curved pattern of epicenters in grid NE and corner pattern of epicenters in grid SW are artifacts of the grid search portion of the beamforming method.](image)
one cluster grid N during falling tide, one cluster grid S during falling tide, and one cluster grid S during rising tide.

Seismic events which happen in the grid N cluster in the extensional zone mostly occur during falling tide. Fast flowing ice moves away from the grounding line as suggested by Barroul (2013; Fig B.7). This velocity gradient causes crevasses to form.

Figure B.13. Cartoon showing the theory of forces at work for each cluster. a-c) outline relevant zone and cluster. Black line is MOA grounding line, green line is ASAID hydrostatic line (Bindschadler, 2011). d-f) show forces at work and location of crevasses during rising tide, g-i) show forces and crevasses at falling tide. Extensional zone cluster and compressional zone rising tide cluster are likely due to horizontal flexure, as evidenced by the velocity of stations G4 in comparison to downstream stations. Events occurring in the compressional zone during falling tide are likely from vertical stresses between ice stuck near Mt. Hope and freely floating ice, demonstrated by elevation difference of G1 and G3.
where the floating ice pulls at the grounding line (Fig B.13.g). These seismic events in the grid N cluster are thus consistent with previous understanding of the extensional force of falling tide.

The grid S cluster present during falling tide lies in the compressional zone. Here the difference in elevation between G1 and G3 suggests G1 undergoes more fluctuation in elevation than G3 (Fig B.8). The ice nearer Mt. Hope moves less, while ice further out is less rigid and more subject to tidal forcing. During falling tide, water moves from under the shelf, which will then experience a decrease in elevation. Ice more strongly buttressed near Mt. Hope will sink less, creating a topographic gradient where ice undergoes extension, promoting crevasses formation (Fig B.13.h). This is also consistent with previous understanding of crevasse formation as tensile fracture on a slope.

Rising tide corresponds to a compressional force, and so the events in the S cluster during rising tide require more explanation. During rising tide, the extensional zone on the outside of the velocity curve is compressed. Ice in the extensional zone then imparts less pressure on the compressional zone. With the force pushing ice from the compressional zone into Mt. Hope lessened, this area undergoes extension similar to what would be expected during falling tide (Fig B.13.f). This pattern of extension and compression is reflected in GPS data which shows that stations G2 and G4 are furthest from each other at rising tide, as a consequence of the velocity difference (Fig B.7). The movement of GPS stations implies that ice within two kilometers downstream of the grounding line moves faster than ice five kilometers downstream at falling tide. Ice further downstream is being compressed by the forces of rising tide, but ice near the
grounding line is able to move faster.

B.4.2 Conclusions

The Beardmore Glacier’s topographical features, in connection with the sharp curve the ice has to take as it outets onto the Ross Ice Shelf, makes it unique from glaciers with a small or nonexistent ice shelf. This is likely why the rates and locations of seismicity were different from what previous work suggests. The Beardmore Glacier moves ice over the grounding line at a pace comparable to other major outlet glaciers on the Ross Ice Shelf (Marsh and others, 2013; Stearns, 2011), and slower glaciers as well as those attached to smaller ice shelves are less likely to divide the glacier into extensional and compressional zones. Continued study of seismic activity near the grounding line will further constrain the relationship between ocean forces, size of ice shelf, and crevasse formation and propagation. This will yield better models of the buttressing effect and thus of Antarctica’s contribution to sea level rise, as well as provide a new way to look at rift propagation and thus calving and ice shelf splintering.

In addition, this research presents the first successful use of the beamforming method for locating seismic events on a glacier. This technique successfully located epicenters from a large data set containing events with emergent arrivals. Although the area within which we found well-located events was small compared to the entire glacier, this area contained spatial patterns of event epicenters that were unexpected, and shed light on the complex interplay between glacier flow, grounding line, buttressing effect, and tidal interactions.
CHAPTER III

C. CONCLUDING REMARKS

This study presents a new way to use beamforming for cryoseismology. Beamforming is successful with emergent, small amplitude seismic signals. From Figure B.5 we can see that the area of “well-located” events for our parameters corresponds to the areas in which the grid N and grid S cluster. Some results were in line with what would be expected from the simplified idea of falling tide as an extensional force, and rising tide creating compression. The general tidal modulation in timing of seismic events and that most seismic events occurred during falling tide agree with previous research (Barroul, 2013). The tidal modulation of GPS station velocities indicate ice downstream of the grounding line moves faster at falling tide and slower at rising tide. Ice nearer areas of high buttressing experiencing less elevation change with tide suggest areas between ice in hydrostatic equilibrium and ice not in hydrostatic equilibrium can be an area of stress and deformation.

However, some of our results indicate there are forces at work which need to be added into this simplified idea. Nearly half the seismic events occurred during rising tide, and ice nearest the grounding line experiencing the largest fluctuations in horizontal velocity, were counterintuitive. We propose that the reason for these results which disagree with the limited previous research are due to the structural constraints of a glacier on a large ice shelf with outlet flow at an angle to ice shelf flow.

The structural constraints imposed by flowing ice bending around an “elbow” creates a sharp turn the ice must follow (Fig B.12), and this quick change causes the volume of
ice on the outside of the elbow to compress the ice on the inside of the elbow.

Events in the grid N cluster, which occur almost exclusively during falling tide, are attributed to extension. With less ocean water pressing against the ice shelf, velocity rates increase downstream of the grounding line and this velocity gradient causes crevassing. The grid S cluster on falling tide is due to extensional forces from the slope created by differences in hydrostatic equilibrium.

If the cluster present during rising tide is crevasse formation, then it must be due to extensional forces. We explain this by calling upon a force that compresses this section, and releases the pressure during rising tide. Looking at the Beardmore Glacier in the context of the Ross Ice Shelf reveals ice takes a sharp turn when it meets ice moving from upstream on the ice shelf. Combining this curve with the buttressing effect of Mt. Hope, located inside the curve, gives us reason to infer this as the mechanism of the compressive force. This force is released during rising tide, when the ice creating this pressure is compressed.

Data from two GPS stations downstream from each other in the extensional zone would lend useful information to this theory. If their motion indicated significantly reduced ice velocity during rising tide, with the station nearer the grounding line experiencing less fluctuation in horizontal velocity than the GPS station further downstream, it would agree with our theory. A broader set of subarray to use for the beamforming method would locate seismic events at other locations. If these improvements were considered for a future study on another glacier which outlets into a large ice shelf at an angle to overall ice shelf flow, results could be tested against our
own. This could determine if either different forces are at work on the Beardmore Glacier than we have theorized, or possibly that the geography of Mt. Hope in relation to the curve of ice plays the major role. In addition, more seismic studies on glaciers that outlet ice directly to the ocean with no turns, or onto smaller ice shelves, should be conducted to see if only falling tide causes events.
REFERENCES


