Analysis of Marginal Ice Zone Noise Events

by

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Abstract

Acoustic transients observed in noise time series, called noise events, are believed to be the major contributor to Arctic Ocean ambient noise. Events are caused by ice fracture processes, each radiating briefly in time. This thesis investigates temporal and spatial distributions of noise events and their dominant source mechanisms – a microlevel approach in the study of Arctic ambient noise.

Underwater ambient noise data from Marginal Ice Zone Experiment 1984 (MIZEX84), at frequencies between 20 and 2000 Hz, are studied. Results show that temporal and spatial distribution of events are characterized by clustering and seem to depend on environmental conditions such as swells and ice concentration.

Causes of Arctic ambient noise are related to the mechanics and breakup of ice. At low frequencies between 10 and 100 Hz, the unloading model of a deformed ice sheet, which occurs immediate the sheet breaks, is shown to be a likely source mechanism. This type of motion includes ridge breaking and surface wave-induced floe breaking. At intermediate frequencies between 100 and 2000 Hz, the ice shear fracture model (caused by surface gravity waves, wind, current, floe shear, and collision) is shown to be the most plausible mechanism, and is thus the main contributor to ambient noise at these frequencies. A stress drop across the fault plane, based on the shear fracture model, is 3 to 4 orders of magnitude smaller than ice shear strength. This low stress drop is probably caused by fault locking, or the non-uniform distribution of the stress release on the whole fault plane, which is similar to rock fracture observed in seismology.

Thesis supervisor: Dr. Ira Dyer

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Contents

Abstract	2
Acknowledgments	3
Table of Contents	4

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謹將此論文獻給我最親愛的爸爸	12
1. Introduction	12
1.1. Background and Motivation	12
1.2. Summary and Contributions	18
1.3. Organization and Data Information	20
1.3.1. The Marginal Ice Zone	20
1.3.2. Environmental Conditions on May 197	22
1.3.3. Ice Motion During May 194 through May 197	24
1.3.4. Acoustic Data	25
2. Methodology	33
2.1. Overview	33
2.2. Position Estimation and Filtering	35
2.3. Time Delay Calculation	36
2.4. Tracking Results	37
3. Event Identification Results of Data Analysis	47
3.1. Event Statistics	47
3.2. Event Spectra	48
3.3. Characterizing in Space and in Time	49
3.4. Event Energy	52
4. Source Strengths of Proposed Ice Mechanisms and Acoustical Mod-	
els	71
4.1. The Ridge-Unloading Model	72

Contents

Abstract	2
Acknowledgments	3
Table of Contents	6
List of Figures	8
List of Tables	11
1 Introduction	12
1.1 Objectives and Thesis Outline	12
1.2 Previous Research on Arctic Ocean Ambient Noise	13
1.3 Summary and Contributions	18
1.4 Background and Data Information	20
1.4.1 The Marginal Ice Zone	20
1.4.2 Environmental Conditions on Jday 167	23
1.4.3 Ice Motion During Jday 164 through Jday 167	24
1.4.4 Acoustic Data	25
2 Methodology	33
2.1 Overview	33
2.2 Spectrum Estimation and Filtering	35
2.3 Time Delay Calculation	36
2.4 Locating Events	37
3 Event Statistics: Results of Data Analysis	47
3.1 Event Statistics	47
3.2 Event Spectrum	48
3.3 Clustering in Space and in Time	49
3.4 Event Density	52
4 Source Strength: Proposed Ice Mechanisms and Acoustical Models	71
4.1 The Ridge-Unloading Model	72

4.1.1	Ridge Breaking	74
4.1.2	Equivalent Acoustic Source: Vertical Dipole Model	79
4.1.3	Dipole Force Data	81
4.1.4	Signature Segregation	82
4.1.5	Dipole Force vs. Decay Rate of Type II & III	83
4.1.6	Discussion on Dyer's Modified Ridge-Unloading Model	83
4.2	Wave-Induced Floe Breakup Model	84
4.3	Floe Shear and Collision Model	88
4.3.1	Floe Collision	89
4.3.2	Floe Shear	92
4.4	Shear Fracture Model	93
4.4.1	Octopole Model and Source Strength	95
4.4.2	Effect of Neglect of Ice in the Octopole Model	99
4.4.3	Propagating Rupture	100
4.4.4	Discussion of the Shear Fracture Model	104
5	Conclusions	134
5.1	Event Statistics	134
5.2	Source Mechanisms	136
5.2.1	Unloading motion (10 - 100 Hz)	136
5.2.2	Shear Fracture Model (100 Hz and above)	137
5.3	Recommendations	138
	Bibliography	140
	Appendices	147
A	Data Analysis	147
A.1	Digitized Data Format	147
A.2	Waterfall Plot Technique	148
A.3	Time Delay Error Estimation	148
A.4	Range Error Estimation	152
A.5	Lloyd Mirror Effect Correction	154
5.1	Cluster Locations in the Second Time Segment	55
5.2	Cluster Locations in the Third Time Segment	56
5.3	Cluster Locations in the Fourth Time Segment	57
5.4	Cluster Locations for Original Deployment during May 1964	58
5.5	Cluster Locations Observed in Event Locations of Clusters in the	59
5.6	Cluster Locations Observed in Event Locations of Clusters in the	60
5.7	Cluster Locations Observed in Event Locations of Clusters in the	61
5.8	Cluster Locations Observed in Event Locations of Clusters in the	62
5.9	Cluster Locations Observed in Event Locations of Clusters in the	63
5.10	Cluster Locations Observed in Event Locations of Clusters in the	64
5.11	Cluster Locations Observed in Event Locations of Clusters in the	65
5.12	Cluster Locations Observed in Event Locations of Clusters in the	66

List of Figures

1.1	Location of MIZEX84 and the Drift of Kvitbjorn and Polar Queen .	28
1.2	Shear Stress and Shear Moment due to Wind and Current Stress. Data provided by N. Makris [1]	29
1.3	Sound Speed Profile for MIZEX84	30
1.4	Apex Array Configuration When First Deployed on J164	31
1.5	Ambient Noise Time Series at 25, 50, 100, 300 and 1000 Hz	32
2.1	Procedure of event analysis	40
2.2	Waterfall plot for event detection. Each trace in a bundle represents a different channel	41
2.3	Six Signature Types	42
2.4	Event frequency range determination	43
2.5	Sine Square Tapered End Boxcar Window	44
2.6	Parks & McClellan FIR linear phase filter	45
2.7	Propagation path	46
3.1	Event Spectrum of Type I	54
3.2	Histogram of Peak Frequency of Events (in 100 Hz bin) within Each Time Segment	55
3.3	Histogram of Event Inter-Arrival Time (in 10 s bin) within Each Time Segment	56
3.4	Histogram of Cluster Durations (in 1 s bin) and Inter-Arrival Time (in 10 s bin) within Each Time Segment	57
3.5	Event Locations in the First Time Segment	58
3.6	Event Locations in the Second Time Segment	59
3.7	Event Locations in the Third Time Segment	60
3.8	Event Locations in the Fourth Time Segment	61
3.9	Phone Locations for Original Deployment during Jday 164.	62
3.10	Straight Line Pattern Observed in Event Locations of Clusters in the First Segment(a)	63
3.11	Events Cluster in Space Observed in the First Segment	64
3.12	Straight Line Pattern Observed in Event Locations of Three Clusters, which inter-arrive at swell periods, in the First Segment(b)	65
3.13	Curve Observed in Event Locations of Clusters in the Second Seg- ment	66

3.14	Curve Observed in Event Locations of Clusters in the Third Segment	67
3.15	Effective Detection Area of Arrays	68
3.16	Event Density for All Time Segments	69
3.17	Ambient Noise Level Measured in Various Frequency Ranges (25-50, 90-110, 290-310, 980-1000 Hz)	70
4.1	Time Series of a multi-pulse event which consists of 3 consecutive pulses of decreasing peak frequencies, namely 345, 125, and 60 Hz .	108
4.2	The spectrogram of Phone D3 of the event in Figure 4.1	109
4.3	Geometry of a rigid circular plate tilting due to a force couple	110
4.4	The ratio of radial to tangential stress vs. r/l	111
4.5	A Vertical Dipole Model At the Sea Surface	112
4.6	Normalized Pressure Field of a Vertical Dipole vs Horizontal Range at two frequencies, namely 100 and 1000 Hz	113
4.7	Dipole Force vs. Event Peak Frequency within Each Time Segment .	114
4.8	Dipole Force vs. Event Peak Frequency within all Time Segments. .	115
4.9	Dipole Force vs. Peak Frequency of Signature Type I to VI. The dotted line is a least-square fit.	116
4.10	Dipole Force vs. Peak Frequency of Type I Events. The dashed line is extrapolated from the central Arctic data for $f < 80$ Hz	117
4.11	Dipole Force vs. Decay Rate of Type II and III	118
4.12	Dipole Forces of Type I events from the MIZ and the Central Arctic compare with Ridge-unloading Model	119
4.13	Strain at mid-floe normalized by Surface Gravity Wave Magnitude versus Floe Length for Three Different Wave Periods, 12, 15, and 17 seconds.	120
4.14	Floe Displacement before and after Breakup with wave period of 12 sec.(a) floe length = 100m,(b) 200m, and (c) 400m	121
4.15	Floe Length versus Peak Frequency of Unloading Motion When an Ice Floe Breaks.	122
4.16	Nine Types of Quadrupoles or Force Couples(Kim89). The axial arrangements are identical, as are the transverse (i,j); coordinate changes in each of these two groups distinguish one from another. .	123
4.17	Spherical Coordinate System for the Octopole Model. The free sur- face is at $z = 0$	124
4.18	Double Couple for Group I (Lateral Octopole I)	124
4.19	Two Possible Double Couples for Group II (Lateral Octopole II) . .	125
4.20	Lat1 Octopole Source Strength vs. Peak Frequency of Type I Events	126
4.21	Lat2 Octopole Source Strength vs. Peak Frequency of Type I Events	127
4.22	Lat1 Octopole Source Strength vs. Peak Frequency of Type II & III Events	128
4.23	Lat2 Octopole Source Strength vs. Peak Frequency of Type II & III Events	129
4.24	Comparison of Kim's result and octopole model	130

4.25	Nulls of sinc function observed in event spectra	131
4.26	(a) Event and receiver locations of spectra shown in Fig. 4.25. The circles are the receivers, the cross is the event location, and the arrow is the hypothesized source orientation. (b) Time series of the event in Fig. 4.25 and 4.26(a). The frequency range is 60 to 1000 Hz	132
4.27	σ_B reduced from $S_{octo,lat1}$ in Fig. 4.20 vs. Peak Frequency	133
A.1	Filters for digital data acquisition system: (a) low-pass, and (b) high-pass	155
A.2	Digital Tape Format	156
A.3	Time Delay Processor[30]	156
A.4	Example # 1	157
A.5	Time Series of Example #3	157
A.6	The Configuration and Force Equilibrium Diagram of Drift Phone Cable	158
A.7	Geometry of the source and its image, and the receiver	158

List of Tables

1.1	Summary of ice composition and floe size distribution from a submarine expedition by Kozo and Tucker, during the Winter of 1974 across the Denmark Strait between 67N and 70N	21
1.2	Summary of hydrophone conditions of MIZEX84	26
1.3	Information for the four time segments	27
3.1	Population of event types during time segments	48
3.2	Cluster statistics	50
4.1	Proposed ice mechanisms and their characteristics (Extremely low frequency band: < 10 Hz; low: < 100 Hz; mid: 100 to 1000 Hz; high: > 1 kHz).	73
4.2	Critical strain for various crack sizes	85
4.3	The minimum surface wave magnitude needed to cause additional fracture in an ice floe of various length (400 m, 200 m, and 100 m) with various crack sizes (1mm, 10cm, and 1m) at various wave periods (12 s, 15 s, and 17 s).	86
4.4	Summary of characteristics of octopoles	97
4.5	Calculation of null spacings and the source orientation	103
4.6	Summary of all $\langle B(\phi) \rangle$	105
A.1	Digitization information for the four segments	148
A.2	Phone information for Example #3	151
A.3	Current speed and cable excursion	153

Chapter 1

Introduction

1.1 Objectives and Thesis Outline

Individual acoustic transients observed in noise time series, called acoustic noise events, are believed to be the major contributor to Arctic Ocean ambient noise ([47,55,8,15,35]). This thesis aims to understand better the temporal and spatial distributions of acoustic events extracted from Marginal Ice Zone Experiment 1984 (MIZEX84) data and the dominant source mechanisms causing the noise events.

The data studied in this thesis are reduced and interpreted as follows:

- The event locations, spectra and signature types are determined. Event densities for different time periods are calculated.
- The source strengths of the events are obtained by adopting acoustical models, namely the dipole and octopole. Ice mechanisms are discussed and examined as sources of the events.

The thesis is organized in this manner: Background information, such as basic environmental conditions and the acoustic data analyzed, are presented in this chapter. In Chapter 2, I show the methodology of event study up to locating the events. The results of event studies such as temporal and spatial distributions, and density of events are presented in Chapter 3. Chapter 4 discusses the proposed ice

mechanisms and their equivalent acoustic models. Then source strength inverted from the proposed acoustic model is presented. Chapter 5 contains the major results of this thesis, conclusions, and recommendations for future research. Appendices include the data format, waterfall plot technique, errors related to event location, and the correction for the Lloyd mirror effect.

1.2 Previous Research on Arctic Ocean Ambient Noise

Research on Arctic ambient noise can be divided into two major areas: one is on environmental correlation with ambient noise levels in various frequency ranges, and the other is on event physics which involves theoretical modeling and data interpretation to extract source parameters. Although my work belongs to the latter category, the former serves a supportive and guiding role. Previous research is reviewed according to its focus in the two areas mentioned above.

1. Study of environmental correlation with Arctic ocean ambient noise:

Milne and Ganton [47] presented results of three experiments in three different seasons of the year, late Spring, Summer and Winter, for shore-fast ice. They showed ambient noise spectra at different times, reported on the quality of the noise time series, and speculated that thermal ice tensile cracks to be the source of the ambient noise for both Spring and Winter, and ice floe relative motion as the cause during Summer. Yet they did not present enough evidence to verify their speculations until later. Ganton and Milne [21] explained the cause of under ice ambient noise between 100 Hz and 10 kHz, again for shore-fast ice. They examined the Winter data of the previous paper and reported different behavior for the ambient noise for two different frequency ranges. From 200 Hz to 800 Hz, the data showed highly impulsive noise, and the noise level went down as the air temperature rose, and the noise went up as

the air temperature dropped. From 1 kHz to 10 kHz, the data showed the noise to be Gaussian (non-impulsive) and also appeared to depend on wind speed. Therefore, they concluded that thermal ice tensile cracks caused the 200-800 Hz band noise and wind-related forcing for the 1 kHz to 10 kHz band. The above studies by Milne et al. were mainly in shallow water.

Green and Buck [22] correlated noise level with wind speed for data collected in deep water for the central Arctic pack ice when ice started to break up in the Spring. Their work and some others on ambient noise correlation with wind speed were reviewed and summarized in both Shepard's [55] and Y.M. Chen's theses [8].

Makris and Dyer [39]. showed that long term variations (greater than 1 hour) of low frequency ambient noise (peak at 15 Hz) was highly correlated with composite measures of stress applied to the ice by wind, current, Coriolis forces, and sea surface tilt.

All the foregoing were for pack ice or shore fast ice. Work seeking environmental correlates for the MIZ ambient noise are few. Diachok et al. [12] showed the spatial variability of ambient noise across an ice edge. Ambient noise levels at central frequencies of 100, 315, and 1000 Hz had maximum values at the ice edge compared with levels in the open water and well into the ice pack. The ice edge levels seemed dependent upon the period of incident surface gravity waves. Diachok and others [18] in later experiments at the ice edge showed that the ice edge contains 'hot spots' which seemed to relate to positions of ocean eddies.

2. Study of source mechanisms, event physics:

Milne pioneered in source mechanism studies. He considered thermal ice tensile cracks as the noise source mechanism and computed the air temperature cycle's effect on the ice sheet [46]. He also presented a very satisfactory result

showing that air temperature correlated well with noise level in 150-300 Hz band. He calculated the rate of temperature change needed for ice to reach its fracture strength with both an elastic and visco-elastic model. However, I find the ice thermal expansion coefficient he used inappropriate in view of more recent data (Refer to Cox [10]). In his earlier work, Milne [45] proposed a statistical model for Arctic ambient noise as an accumulation of the ice crack noise. The model assumed that ice crack noise events are a number of independent identical random pulses. This assumption is not correct since ice events of the same kind as Milne described vary in strength with frequency (Refer to Townsend-Manning [60] and this thesis). In another paper, Milne et al. [48] suggested that the noise-generating mechanism may vary with the region in the Arctic Ocean, as was shown in other experiments.

Substantial progress was made by Lewis et al. [35,36] concerning thermal tensile cracking of ice. They concluded that heat flux, rather than air temperature, is important in causing ice fracture under thermal tension [35]. Using the heat balance as input, and considering ice as visco-elastic, they calculated the ice stress due to the thermal effect [36].

Ice tensile cracks due to thermal expansion are not a major issue in this thesis, as in the foregoing studies. For shore-fast or pack ice conditions, thermal cracking is important during the Spring and Winter [21,35]. Since MIZEX84 was conducted in the Summer, this mechanism is an unlikely source of the events analyzed in this thesis.

Pritchard [53] proposed that ambient noise is mainly caused by the energy dissipated during ice ridge building. Events caused by ridging are considered absent in the MIZ.

For pack ice, Shepard [55] suggested that low frequency ambient noise (below 80 Hz) level was the sum of short duration bursts of sound due to ice motion.

He presented horizontal and vertical directional analyses to show that in quiet periods the noise originated from particular azimuthal and radial sectors where many transient peaks were present. During noisy periods the noise field was isotropic azimuthally and radially. From the directional properties of the receiving array he also concluded that the noise source must be at the surface, i.e. the ice cover.

Dyer [15] presented a more general picture of the physics of noise events. He discussed possible event mechanisms in the low and mid-frequency portions of the noise frequency spectra. He proposed a slip model for ice fracture. I will use this slip model to represent shear fractures of ice floes (Sec. 4.4). He also proposed two ice models based on the results of Makris and Dyer (mentioned in the previous paragraph) [16]: ‘ridge-unloading model’ and ‘ice beam rafting model’. I will use his ridge-unloading model as one of my source models (Refer to Sec. 4.1).

The theoretical modeling of ice fracture as a source radiating acoustic energy in water was done first by Stein [59]. He modeled ice tensile cracking by placing a monopole inside the ice plate and derived the radiated underwater acoustic pressure field. He also showed agreement between wave modes excited by the monopole and events observed in a lake experiment. Langley argued that Stein’s monopole model of an ice tensile crack was not right since the crack model must include a quadrupole as well [33,34,32]. Kim [29] used a full-wave numerical tool instead of analytical means to solve for radiation from ice motions due to seismic moment. He adopted seismic fault plane and force couple models. For the tensile crack, he superposed an in-line force couple and a body force that radiated in all directions, the latter equivalent to the monopole used by Stein. He also dealt with shearing and grinding in different directions at different plane orientations. I will compare some of his results with the simplified octopole model I use for shear fractures. He

computed radiation patterns of propagating fractures and showed that the event spectrum would be multiplied by a sinc function, as shown by Aki [2].

Farmer et al. studied ice events collected from shore-fast ice in shallow water [19]. They computed the acoustic radiation by placing a monopole in a fluid layer over another fluid layer. Although ice is poorly modeled as a fluid, they advanced further than Stein by taking crack length into consideration. They integrated the solution for a single monopole over an assumed crack length and found agreement between measured event spectral components and the theory. (Since they collected the data at shallow water, they had to resolve bottom reflection and multipath arrivals)

Dyer [17] and later Keenan [25,26,27] studied events through categorizing them according to signature types. Townsend-Manning [60] was the first to count events and to study their temporal and spatial distributions with a massive data base. She analyzed 199 events extracted from 580 minutes of central Arctic pack ice data. The event data were bandpass-filtered within 20 to 80 Hz. Using a vertical dipole to represent the source and its image, she computed the source strength. I will use this dipole model as well and compare her results with mine. My data are from the MIZ while hers were from the central Arctic, and the frequency range of my data is from 20 Hz to 2000 Hz.

In my earlier work with Buckingham [7], we used a monopole in a sphere to model the floe bumping. The signature of the proposed model matched with the observed event signature, i.e. repeated pulses of identical time intervals and pulse shapes. The pulse repetition was due to the spherical boundary reflection. However, no further data were analyzed to see if the model is valid. Section 4.3.1 discusses this model in detail.

1.3 Summary and Contributions

The major results are summarized as follows:

1. A total of 458 events, in four 40-minute time segments, are analyzed over a region of about 1 km^2 in the MIZ. They are detected at frequencies between 20 and 2000 Hz. Six event signature types are identified. The signature types observed in the MIZ are the same as in the central Arctic.
2. Events cluster in time and space. One third of the event clusters at the first 3 time segments recur at periods correlated well with the dominant swell periods in the MIZ, which are between 10 to 18 seconds. The cluster inter-arrival time does not correlate with the swell periods at the 4th segment, showing that swells are important environmental forces at the first 3 segments but not at the 4th segment. The sea state measured confirms this finding – the swell reaches its maximum at the time between 1800Z on J166 and 0400Z on J167 (the first 3 segments are within this time period) and then dies down. Events cluster in space in the first 3 segments and distribute more evenly in the 4th segment. The differences in the temporal and spatial distributions of events between the first 3 segments and the last may be due to changes in environmental forces causing the events or changes in environmental conditions, such as swells which peak during the first 3 segments and then dies down, and the ice concentration which fluctuates from the 1st to the 4th segment. These two environmental factors seem to dominate the event distributions.
3. The chronological order and coherent locations of events in some clusters during the 1st segment, which inter-arrive at swell periods, suggest crack propagation. The cluster locations at the 2nd and 3rd segments even match the hypothesized floe boundary, which suggest floe interactions as causes of the events.

4. The event density varies with time, up to 10 times between the first and last time segment. The maximum event density is 0.58 per km^2 per hour per Hz at 300 Hz in the first segment. Event densities increase as f at frequencies below 200 Hz, decrease as f^{-1} at frequencies above 300 Hz, and maximize at frequencies between 200 and 300 Hz. The change in trend may indicate a transition in the source mechanism at frequencies below 200 Hz and above 300 Hz.
5. Low-frequency pulses (e.g. 60 Hz) occur immediately after mid-frequency pulses (e.g. 125 and 345 Hz), confirming that shear fractures (source mechanism at frequencies above 100 Hz) are the pre-cursors to the unloading motion of a deformed plate which occurs right after the plate breaks (source mechanism at frequencies between 10 and 100 Hz).
6. Shear fractures are the most probable source mechanism at frequencies above 100 Hz, which are the breaking processes of the ice due to surface gravity waves, wind, current, and floe inter-actions. Transformed by a double couple and its image, an octopole is used to represent the shear fracture at a fault plane. The moment of the double couple ranges between 10^2 and $10^5 N\cdot m$; its frequency dependence is $f^{-2.5}$. The stress drop of the fracture is 3 to 4 orders of magnitude smaller than the shear strength of the ice. This low stress drop, also observed in seismology, may be the result of fault locking, or the non-uniform distribution of the stress release on the whole fault plane, caused by asperities on the fault plane. Also, the nulls observed in the event spectra indicate fracture propagation.
7. The unloading motion of a deformed ice sheet, which occurs immediately the ice breaks, is shown to be a likely source mechanism at frequencies below 100 Hz. Many fractures and cracks in the ice precede this mechanism. A vertical dipole is used to represent the volume change in the unloading process. The dipole forces of some of the more important events agree well with the ex-

trapolated values from the previous results from the central Arctic. However, the ice stresses corresponding to the dipole forces are far less than the ice strength. The creep behavior of the ice at low strain rate may be the cause of this low stress result [14].

1.4 Background and Data Information

This section includes a general description of the MIZ environment, environmental conditions and the ice floe motions on Jday 167, as well as the acoustic data analyzed.

1.4.1 The Marginal Ice Zone

The marginal ice zone (MIZ) is the region where Arctic pack ice meets the open ocean; it is characterized by a gradient of ice concentration. Due to the complex nature of ice dynamics and their great influence on the atmosphere, the MIZ plays a vital role in the understanding of noise generation, air/ice/water interactions, wave propagation, etc.

A series of Marginal Ice Zone experiments – MIZEX83, MIZEX West, MIZEX84, MIZEX87, and CEAREX – were carried out with the cooperation of scientists from several countries. The data analyzed in this thesis are taken from MIZEX84. The overall goal of MIZEX84 is to understand further the mesoscale processes that dominate the motion and characteristics of the ice margin. Within the focus of the exercise, various areas of interest are oceanography, ice dynamics, meteorology, remote sensing, and acoustics. MIZEX84 encompassed the ice edge of the Fram Strait, between 81N and 79N in latitude and 10W and 10E in longitude (Figure 1.1). The ice edge extended from 80N,10E to 76N,10W with an overall NE-SW orientation.

The environment of the MIZ can be understood from several perspectives [62,24,50,37,49]

1. Ice Cover: This includes floe size distribution, ice composition, and ice com-

Description	Regime 1	Regime 2	Regime 3	Regime 4	Regime 5
Distance from ice edge	0-58km	58-94km	108-153 km	153 km-166km	above 166 km
Floe diameter range	3-50m	50-75m	50-100m	100-150m	over 400m
Ridge occurrence & draft (d)	none	$d \geq 5\text{m}$	increased	$d \simeq 12\text{m}$	central Arctic

Table 1.1: Summary of ice composition and floe size distribution from a submarine expedition by Kozo and Tucker, during the Winter of 1974 across the Denmark Strait between 67N and 70N. The season and the area are very different from MIZEX84, and thus this table is at best crudely suggestive of MIZEX84 conditions.

pactness. Ice banding is also observed in the experiment. Kozo and Tucker [31] identified five regimes of ice as they were transiting inwardly from the ice edge. Their observations are summarized in Table 1.1, which represents a typical ice profile of the MIZ.

2. Oceanography and Ice Edge: Cold water meets warm water at an ice front. Northerly driven winds blow ice into the open water, causing a diffuse ice edge. Southerly winds cause a compact ice edge. In the absence of wind, the radiation pressure of waves will maintain a compact ice edge.
3. Oceanography and Heat Flux: The upper mixed layer of the ocean is affected by salt injection into the water from ice formation during the Winter, and by injection of fresh water from melt during the Summer. The cooling and heating effects of the atmosphere on the water is balanced by the ice; the direction of heat flux changes with the season and with temporal effects such as solar insulation and cloud cover.
4. Internal Stresses and Ice Drift: Due to the discontinuous nature of the ice floes in the MIZ, internal stress does not build up in the same manner as in

the central Arctic where ice sheets are more continuous. Of course, internal stresses can be transmitted via floe collision. Free ice drift results in a small to zero internal stress in the ice floes.

5. Meteorology: A strong thermal gradient in the sea surface temperature is present across the ice edge, which results in varying the atmospheric boundary layer height from 300 m over the ice to 700 m over the open water. This also introduces a steady baroclinic wind of 5 to 7 m/s. The cloud pattern differs in regions above the ice and the open water, e.g., fog over the ice and low clouds over open water. Wind stress on the ice is an important driving force.

6. Other Oceanographic Properties:

- Ice/ocean eddies exist in the MIZ. The eddy generation mechanisms includes both baroclinic and barotropic instabilities, topographic inhomogeneity, wind-driven upwelling along the ice edge, and the interaction between the ice edge and existing eddies. Typical features of eddies are ice bands with cyclonic rotation and higher ice concentration in the center. Temperature and salinity have strong horizontal gradients within eddies, such that the sound-speed structure changes more than 15 m/s. Ice melt and ice-edge retreat are greatly influenced by eddies. Eddies transport warm water into and ice floes out of the ice cover. The warm Atlantic water, in association with eddy motion, can melt ice at a rate 10 times faster than without eddy motion. The ice-edge also retreats faster due to the increased bottom ablation associated with eddies. High ambient noise levels at the site of eddies were reported by Diachok [18].
- Internal waves are another important feature in the MIZ. The water column in the MIZ is typically stratified, possessing a strong two-layer salinity and density structure caused by ice melting (in the Summer). Internal waves may be generated due to the interaction of ice keels with the oceanic stratification, the response of the stratified ocean to a non-

uniform wind stress over the ice and open water, or interaction with nearby bottom topography. Interactions between internal waves and ice floes are speculated to cause ice convergence, which enhance the floe interactions and thus ambient noise generation.

- The sound speed profile of the MIZ in the Summer is characterized by a sharp surface duct due to a halocline created by ice melting which forms a less dense surface layer. The sound speed also varies significantly in range over several km. Beneath 800 m, the sound speed is dominated by the effect of hydrostatic pressure, giving the usual deep ocean gradient of about 0.016 s^{-1} . Between the surface layer and 800m, the warm West Spitsbergen current flowing northward, which heats up the water column, produces complicated sound speed profiles such as in Figure 1.3.
- An important environmental forcing, current stress is associated with the current speed relative to the ice drift speed and the ice-water drag. In the MIZ, the variations of ice concentration and floe size can cause changes in the ice-water drag.
- Coriolis force and sea surface tilt exert stresses on the ice floes, transmitting into ice motion instead of internal stress due to the discontinuous nature of ice floes. If ice is of a ridged structure, the stresses accumulate at the ridges.

1.4.2 Environmental Conditions on Jday 167

This section presents the environmental conditions during Jday 167, the period emphasized in my acoustics study. These conditions include ice concentration, wind and current stresses, sea state, and sound speed profile.

- Ice concentration: The mean ice concentration [1] is about 50% to 70% on Jday 167. Data suggest that the ice concentration fluctuates during all four time segments studied in this thesis.

- Wind and current stress: Wind and current are expressed through shear stress and shear moment. Shear stress is defined as the vector sum of wind stress and current stress. Shear moment is the difference between the two [1]. Figure 1.2 shows the stress and moment as functions of time. Note that the shear stress stays the same throughout the 1st and 3rd segments but maximizes during the 4th segment. However, shear moment increases from the 1st to the 3rd segment and drops back to the level of the 1st segment during the 4th segment. (Refer to Makris [38] for more details.)
- Sea state: The sea state in the open water region of the MIZ changes during Jday 166 and 167 [1]. It increases at the beginning of Jday 166, reaches a maximum around 2100Z that day, remains at the same level until 0600Z on Jday 167, then decreases and dies down completely at the end of Jday 167. Note that the first 3 time segments of my data reside at the peak of the sea state.
- Sound speed profile: Figure 1.3 shows the sound speed profile used in the data analysis (except in locating events). The sharp gradient at depths above 60 m is a typical feature in the MIZ (surface duct). Between 60 and 360m, the sound speed increases gradually. The sound speed decreases at depths between 360 and 520m and remains constant until the bottom; thus the rays that reach below 360 m suffer due to bottom bounces. The sea bottom at the experiment site is about 800 m.

1.4.3 Ice Motion During Jday 164 through Jday 167

A horizontal hydrophone array was deployed as a cross on an ice floe of 400m diameter on J164. Figure 1.4 shows the array configuration when first deployed. A total of 13 hydrophones (also called drift phones since they drift with ice floes) were deployed. Phones D11, D23, D18, D12 are 12.5m northward, eastward, southward, and westward respectively from the apex, D5. Similarly, D3, D8, D4, D19 are 50m

from D5. Phones D6, D2, D24 and D20 are 200m away from the apex. The north-south orientation of the array is only valid at the time they were deployed, since the floe rotates afterwards.

At 2200Z of J166 the floe drift trace shows that the big floe has broken up. At this moment, the original floe has broken into at least six pieces. The 200m ring has completely broken away, with phones D6, D20, D24, D2 drifting separately. The rest of the nine phones were divided into two floes, with D4, D8, D19 staying together and D3, D11, D12, D18, D23, D5 in another group. This is true through J167. After J167, D3 is missing, probably since the floe on which the five phones are attached breaks.

1.4.4 Acoustic Data

Acoustic data are acquired by deploying hydrophones through ice floes adjacent to an ice-strengthened ship, Kvitbjorn. Four different types of hydrophones are utilized (See Table 1.2). The D-series consist of phones deployed from drifting floes. A sensor tracking system (STS) is used to locate each phone [11]. Only data acquired by D-phones are analyzed in this thesis. Since the phones are far away from the ship, ship noise is absent, especially since the ship is completely shut down during ambient noise runs except for a well-isolated deck-mounted motor generator. The depth of the D-phones is 60m. Acoustic experiments include ambient noise, refraction, reflection, scattering, and transmission experiments.

The acoustic data analyzed in this thesis are digitized from analog tapes which recorded twelve channels (D2, D3, D4, D5, D6, D8, D10, D11, D12, D18, D19, D20). D3, D11, D12, D18 and D5 are named as Array # 1 and D4, D8, D19 Array #2 throughout the rest of the thesis. The dimensions for the arrays are 72.5m by 5m for Array #1 and 200m by 100m for Array #2. The events are detected with these 12 phones. The distance between Array #1 and Array #2 are 500m during the 1st segment, increases to 1km at the 2nd segment, and remains at this distance

Series	Total #	Depth	Locations	Linkage
D	24	60m	remote	telemetry
X	3	60m	same as ship	hard-wired
HF	12	< 10m	same as ship	hard-wired
V	—	various	same as ship	hard-wired

Table 1.2: Summary of hydrophone conditions of MIZEX84

throughout J167. Phone D20 is 200m away from Array #1 on J166 and J167; D2, D24 are 300m away from Array #2 on J166 and J167.

The data analyzed consist of four 40 minute segments spanning 20 hours beginning at 2200Z on J166 and ending at 1800Z on J167. The choice of this time period for detailed data analysis is based on two considerations:

1. The ambient noise levels between 25 and 1000 Hz on J167 are the lowest in the two-week experiment period. Events are easier to detect when the ambient noise level is low [60]. Figure 1.5 shows the ambient noise time series at various frequencies. Each data point of the time series represents the spectral density at the frequency specified averaged over a 20 Hz band.
2. The analog tape recordings pose a limitation on which time period to choose. The time period I choose is the only time when Array #1 & #2 are recorded. After 1800Z on J167, only two phones of Array #1 are recorded.

The J167 data are analyzed for brief periods in three-hour intervals. (The 12-hour interval between the 3rd and 4th segment is due to an interfering transmission and refraction experiment, preventing any meaningful event detection.) In order to obtain a broader frequency range, these data are digitized from analog tapes since the digital tapes also obtained during the experiment contain data only from 1 to 250 Hz. Due to the limitation of the equipment used for digitization, the sampling frequencies are different for the four segments. (See Appendix A.1 for more details.)

Seg. #	Start Time	End Time	Sample Interval	f_{min}	f_{max}
1	166:22:15	166:22:55	0.125 msec	8 Hz	2 kHz
2	167:03:24	167:04:04	0.125 msec	8 Hz	2 kHz
3	167:06:31	167:07:11	0.250 msec	4 Hz	1 kHz
4	167:17:20	167:18:00	0.250 msec	4 Hz	1 kHz

Table 1.3: Information for the four time segments

Table 1.3 summarizes the details of the four segments. f_{min} and f_{max} are the lower and upper frequency limits of the data within each time segment.

Figure 1.1: Location of MIZEX84 and the Drift of Kvitbjorn and Polar Queen

Figure 1.2: Shear Stress and Shear Moment due to Wind and Current Stress. Data provided by N. Makris [1]

Figure 1.3: Sound Speed Profile for MIZEX84

Figure 1.5: Ambient Noise Time Series at 25, 50, 100, 300 and 1000 Hz

Chapter 2

Methodology

2.1 Overview

This chapter presents the methodology for analyzing the data. The procedure of detecting and analyzing an event is shown in Figure 2.1. Preliminary event hunting is based on waterfall plots (see Figure 2.2 and Appendix A.2 for waterfall plot details). Although an acoustical event is detected when the pulse amplitude is above the background noise level, only events detected on three or more channels are counted and analyzed; a minimum of three channels is needed to locate an event. While examining the waterfall plots, I first roughly count the events followed by detailed counting with the event signature enlarged.

Events are classified by six signature types (see Figure 2.3). Type I is called a “pop” which is characterized by the third time derivative of any smooth monotonic change in value. Types II and Type III are both called “damped sinusoids”. The difference between the two is that Type II can be considered as a complex “pop” with one or two more cycles than a Type I “pop”, where Type III is an unmistakable damped sinusoid. Type IV is called a “pure tone” which has a finite sinusoidal waveform. Types V and VI both have series of pulses which individually can be any of the first four waveforms. Type V is distinct from Type VI in that the pulse separation for the former is the same throughout the series, but not so for Type VI.

For example, if two pulses occur successively but the third pulse does not have the same pulse separation as between the first two, I classify the pulses as a Type VI event. (The ambiguity in classifying events lies mainly in distinguishing between single-pulse events (Type I to IV) and multiple-pulse events (Type V and VI), which strictly speaking can be treated as several single-pulse events.) Guidelines used to identify Type V and VI events are:

1. The pulse separations of the series pulses are the same for all channels that detected the pulses; the locations of each individual pulse of the series are the same.
2. The pulse separation is less than the maximum time delay to reach the phones which detected the event. For example, with 0.0433s as the maximum time delay for Array #1, the pulses are considered as one event of either Type V or VI if the pulse separation is less than 0.0433 s. (Since the maximum spatial phone separation is 62.5 meters and the water sound speed is 1440 m/s)
3. The spectral content of all pulses in an event is identical.

There is subjective variation in identifying event signatures; my taxonomy is slightly different from those used by my predecessors (Dyer [17], Keenan [25]). Nevertheless, event classification on the whole will prove to be of importance in reaching physical conclusions.

Time delays and phone positions are needed in locating the events. Time lags are determined by first filtering the events and then cross-correlating them among recorded channels. The frequency range of an event is estimated by comparing the event spectrum with the ambient noise spectrum prior to the event. (See Figure 2.4) Note that filtering the event according to this frequency range will obtain a higher SNR. Then signals from adjacent phones are cross-correlated with the reference signal.

Event locations are determined in two steps. First, computed range and time delay are fitted by a straight line, using the least square method, for each grid point in an area of $2km \times 2km$, with the phone array at the grid center. With an initial guess for the event location, the coupled range-position equations are solved by Taylor-series expansion and iteration, an algorithm used by Duckworth et al. [11] for Sensor-Tracking.

The following sections describe the above calculations (spectrum estimation, time delay estimation, and locating events) in detail.

2.2 Spectrum Estimation and Filtering

Power spectrum estimation by a FFT algorithm is performed for each channel of every event. The power spectrum level in dB re $1 \mu Pa$, and 1 Hz is:

$$10 \times \log S(f) = 20 \times \log |X(f)| - 10 \times \log T + 159 - (\text{Pre_Gain}) + 3 \quad (2.1)$$

where

$S(f)$ = one-sided power spectrum [μPa^2 per Hz],

$X(f)$ = two-sided Fourier Transform of event time series [*Volt* per $\sqrt{Hz}$],

T = time difference between the first arrival and the end of the last arrival of the event [*sec.*],

159 = sensitivity of the hydrophones [*dB* re $\frac{1 \text{ Volt}}{1 \mu Pa}$],

Pre_Gain = pregain setting in data acquisition (30 dB for the drift phones which acquired the data), and

3 dB= for the one-sided spectrum correction.

The time window used is a boxcar with sine squared tapering at two ends. The duration of the window is roughly equal to duration between the first arrival and

the end of the last arrival, *i.e.*,

$$w(t) = \begin{cases} \sin^2\left(\frac{\pi}{2} \frac{t}{0.05T}\right), & 0 \leq t \leq 0.05T \\ 1, & 0.05T \leq t \leq 0.95T \\ 1 - \sin^2\left(\frac{\pi}{2} \frac{t-0.95T}{0.05T}\right), & 0.95T \leq t \leq T \end{cases} \quad (2.2)$$

Figure 2.5 shows the window in non-dimensional time and its spectrum.

The peak frequency of an event, which is considered as the frequency component of the major energy lobe of the event, is determined by comparing its spectrum with the ambient noise spectrum preceding the event (See Figure 2.4).

Once the frequency range of an event is determined, a FIR linear phase filter [51] for the particular frequency range is used to filter the data. Figure 2.6 shows a typical filter and its spectrum.

2.3 Time Delay Calculation

Time delay estimation is the first step of source-locating [30]. First, a reference window is chosen among the event time series of all channels, usually it will be the event time series of the channel in which the event arrives first. Second, cross-correlations of each channel with the data in the reference window are calculated. Third, I choose any one of the channels as the reference channel; its time delay is zero. Fourth, the time difference between the maxima of cross-correlations of each channel and the reference channel is calculated. The time delay is the time difference between the maxima, *i.e.*,

$$D_i = \tau|_{\max(R_{x_i w_{ref}})} - \tau|_{\max(R_{x_{ref} w_{ref}})}$$

where

D_i = time delay of the i th channel,

w_{ref} = reference window,

x_i = time series of the i th channel,

x_{ref} = time series of the reference channel,

R_{xy} = cross-correlation of x and y ,

$\tau|_{max(R_{xy})}$ = time when R_{xy} reaches the absolute maximum.

The error in the calculated time delay is about 5×10^{-3} s in this data set. Detailed calculations of the estimation error are presented in Appendix A.3.

2.4 Locating Events

Time delays, receiver positions, and propagation paths are the three essential inputs in locating events. Time delay estimates have already been discussed in the previous section. The propagation path is assumed to be a direct water path from the source to the receivers in locating all events. (See Figure 2.7) The travel time error is 0.004 s with use of the direct path instead of the refractive path based on the sound speed profile in Figure 1.3. This error is comparable to the time delay estimate error. Errors in both time delay estimates and travel times are included in determining the range error. (See Appendix A.4)

Two methods are used in locating events: the 'sound speed fitting method' [59,60], and the 'iteration of linearized range equation method'. The result of the former method is the trial source position for the latter method.

The 'sound speed fitting method' assumes that acoustic signals travel at a constant sound speed, i.e.

$$\frac{r_i}{t_i} = c, \quad i = 1, \dots, N$$

where

r_i and t_i are range and travel time of receiver i to source, respectively, $c = 1440$ m/sec is the sound speed of water, and N is the total number of receivers which detect the event. Similarly,

$$\frac{r_i}{t_i} = \frac{r_j}{t_j} = \frac{r_i - r_j}{t_i - t_j}$$

Defining,

$$dt_{ij} = t_i - t_j$$

then

$$\frac{r_i - r_j}{dt_{ij}} = c$$

The procedure in determining the source location using the 'sound speed fitting method' is: 1) Assume the source location $\underline{x}$, 2) calculate the range from source to receiver i , and 3) fit a straight line through all data points (r_i, dt_{ij}) . The source location is the $\underline{x}$ which minimizes the least-squared error of the line fitted, provided that the slope is within $c \pm 100$ m/sec. The accuracy of the source location, however, is still unknown.

The 'iteration of linearized range equation method' is a modified version of the sensor-tracking system [11]. That is,

$$c(\underline{t}_0 + \underline{\tau}) = \underline{r} = \underline{r}' + \underline{r}_0 = A(\underline{l}, \underline{x}) \quad (2.3)$$

where

$$\underline{r}_0 = c\underline{t}_0,$$

$$A_i(\underline{l}, \underline{x}) = \sqrt{(l_{xi} - x)^2 + (l_{yi} - y)^2 + d^2},$$

(x, y) = source location,

(l_{xi}, l_{yi}) = the i th receiver position,

$\underline{t}_0$ = travel time to reference receiver,

$\underline{\tau}$ = time delays of each receiver with the reference receiver,

c = sound speed in water (= 1440m/sec), and

d = the depth of the receiver.

Equation 2.3 is solved by expanding $A(\underline{l}, \underline{x})$ using the Taylor expansion, which yields

$$A(\underline{x} + d\underline{x}) = A(\underline{x}) + \nabla A|_{\underline{x}}(d\underline{x}) \quad (2.4)$$

So,

$$d\underline{r} = (\underline{r} - \hat{\underline{r}}) = (\underline{r}_0 + \underline{r}' - \hat{\underline{r}}' - \hat{\underline{r}}_0) = \nabla A|_{\underline{x}}(d\underline{x})$$

where $\underline{r}_{0_{ref}} = \hat{\underline{r}}_{0_{ref}}$ for the reference phone. Then solve for $d\underline{x}$ using

$$d\underline{x} = \{\nabla A|_{\underline{x}}\}^{-1} d\underline{r}$$

and use a new $\underline{x} = \text{old } \underline{x} + d\underline{x}$ in Equation 2.4 to repeat the process. The iteration is continued until the error reaches an acceptable value. The accuracy of the final result is

$$\sigma_x = \sqrt{\{\nabla A|_{\underline{x}}\}^{-1} \sigma_r^2 \{\nabla A|_{\underline{x}}\}^{-T}} \quad (2.5)$$

where

σ_x = root-mean-squared error of source location, and

σ_r = root mean square error of $\underline{r}$ (10 meters).

Appendix A.4 provides detailed calculation of range errors.

Figure 2.1. Procedure of Error Analysis

EVENT STUDY

Figure 2.1: Procedure of Event Analysis

Figure 2.2: Waterfall plot for event detection. Each trace in a bundle represents a different channel. Time increases from left to right and bottom to top.

Type I.: Pop

Type II: Complex Pop

Type III: Damped Sinusoid

Type IV: Pure Tone

Type V: Straf (periodic peaks)

Type VI: aperiodic peaks

Figure 2.3: Six Signature Types

Figure 2.4: Event frequency range determination

Figure 2.5: Sine Square Tapered End Boxcar Window

Figure 2.6: Parks & McClellan FIR linear phase filter

Chapter 3

Event Statistics: Results of Data Analysis

Figure 2.7: Propagation path

Chapter 3

Event Statistics: Results of Data Analysis

This chapter presents the results of analyzed data in terms of statistics of event signature types, characteristic spectra, and event occurrence in time and space. The focus of this chapter is the presentation of results. Physical mechanisms which give rise to the events will be discussed in the following chapter.

3.1 Event Statistics

Table 3.1 shows the statistics of event occurrence and the population of each event type for all time segments¹. (Each event is identified by its shape, according to the taxonomy of Chapter 2.) Two trends are worth noting. First, the occurrence of all events diminishes from 205 to 20 during the time period between the 1st segment and the 4th segment, corresponding to 2200Z on Jday 166 and 1800Z on the next day respectively. Second, event statistics also suggest that most of the events are Type II (complex ‘pop’), Type III (damped sinusoid), and Type VI (aperiodic pulse). This trend is true not only for the total number of events but also appears to be

¹After the completion of data analysis, I realize that all Type IV events classified previously should be identified as Type III events

Type	1st seg.	2nd seg.	3rd seg.	4th seg.	Total #
I	15 (8%)	1 (1%)	5 (5%)	4 (20%)	25 (6%)
II	61 (30%)	11 (8%)	18 (19%)	8 (40%)	98 (21%)
III	37 (18%)	43 (31%)	28 (29%)	2 (10%)	110 (24%)
IV ²	17 (8%)	9 (7%)	13 (14%)	1 (5%)	40 (9%)
V	5 (2%)	15 (10%)	12 (12%)	0 (0%)	32 (7%)
VI	70 (34%)	58 (43%)	20 (21%)	5 (25%)	153 (33%)
Total #	205 (100%)	137 (100%)	96 (100%)	20 (100%)	458 (100%)

Table 3.1: Population of event types during time segments

the case during each time segment as well (except for the 4th Segment which has too few to justify a statistically useful conclusion).

Most events are detected by Array #1, 444 out of 458 total events. (As mentioned in Section 1.4.4, the data are collected by two arrays of hydrophones.) Array #2 detected 9 out of 458. There are 5 events detected by both Array #1 and #2, and 10 events detected by all twelve phones. These 10 events detected are not analyzed since their signatures are a series of Type IV pulses and their locations are much further away than all other events, beyond 1 km, appearing to be due to distant sources. So I suspect that these 10 events are related to the mystery noise reported by Duckworth [13] and Song [58].

3.2 Event Spectrum

A typical spectrum of Type I events is shown in Figure 3.1. The spectrum has characteristics of a well defined peak frequency which represents the event time scale, a positive slope at the low frequency end, and a negative slope at the high frequency end [15].

Event spectra are characterized by the peak frequency which represents the fre-

²After the completion of data analysis, I realize that all Type IV events classified previously should be identified as Type III events

quency component of the event's major energy. The histogram of the peak frequency of the events within each time segment is presented in Figure 3.2. The median peak frequency in each segment is $300(\pm 20)$ Hz, except for the 3rd segment which has median peak frequency of 200 Hz.

The spectra of some events show nulls at various frequencies, with the nulls' positions differing from channel to channel. All of Type I events (25 of them) have nulls that exhibit this behavior and are discussed further in Section 4.4.3.

3.3 Clustering in Space and in Time

The events detected are characterized by clustering both in time and space. The median of event inter-arrival time is less than 10 s for the first 3 time segments, and is 40 s for the last segment. Figure 3.3 shows the histograms of event inter-arrival time for all time segments. Exponential behaviors are observed in the histograms of event inter-arrival time for all time segments.

Martin et al. [40] have shown that the collisions of floes have two cluster types: 1) short-period spikes which recur with the period of ocean swell and 2) long-period acceleration events which are not correlated with the swell. For my data, with the dominant ocean swell periods in the MIZ being between 10 s and 18 s, events are divided into clusters for event inter-arrival time exceeding 9 s (I used 9 s instead of 10 s to include a 10% error), with the hypothesis that event cluster in time is related to the swell. Table 3.2 presents the statistics of clusters in each time segment. One third of the clusters detected during the first 3 segments are with the same periods as the ocean swell (between 10s and 18s), but no cluster inter-arrives at the swell period during the 4th segment. This indicates that the ocean swell could be at least the partial cause for the events during the first 3 segments, but not so for the 4th segment. Figure 3.4 shows the histograms of cluster durations and inter-arrival time. An attempt was made to correlate the cluster recurring pattern with the event types but no correlation was found.

Item	1st seg.	2nd seg.	3rd seg.	4th seg.
# of clusters	56	57	49	13
inter-arrival time between 9 s & 19 s	20(36%)	21(37%)	17(35%)	0(0%)
inter-arrival time above 19 s	36(64%)	36(63%)	32(65%)	13(100%)
mean inter-arrival	37.8 s	39.2 s	50.4 s	177.8 s
mean duration	3.0 s	1.9 s	1.5 s	0.5 s
mean # of events per cluster	4	3	2	1

Table 3.2: Cluster statistics

Event cluster in space is shown by the event locations. Figures 3.5, 3.6, 3.7, and 3.8 show the event locations during the 4 time segments, respectively. Note that the events are concentrated near Array #1 in the first 3 time segments but not during the last segment.

In the 1st time segment, straight-line paths can be drawn for some of the event locations. The time history and event locations provide a nice clue to what happened. Figure 3.9 shows the original phone deployment at Jday 164. The floe where the 13 phones deployed is about 400 m in diameter. Figure 3.10 shows the locations of the phones 2 days later and one event cluster. The cluster duration is 10 s, occurring 9.2 s after the previous event. Two straight lines, path #1 and #2, are drawn according to the locations of this event cluster. It starts at path #1 followed by events occurring along path #2, the arrows indicating the sequence.

Figure 3.11 shows the successive clusters within the next 12 minutes have an inter-arrival time within 19 s, but no particular path is observed in this figure. Figure 3.12 shows the floe marked by the events in Figures 3.10 and 3.11, which are considered to be pre-existing cracks, and three event clusters occurring after the

clusters in Figure 3.11. Cluster 15-1 which occurs 159 s after the previous one, has events closely located near phone D3. Then, 14.5 s later, cluster 15-2 starts along a straight line, which seems to be the extension of path #2 in Figure 3.10. Cluster 15-3 occurs 10.5 s later; most of the events seem to be a continuation of cluster 15-1 and moving away from phone D3.

The four figures, Figure 3.9 to 3.12, suggest that fractures occur on the floe, some of which recur with the swell period. Some have a connected path which indicates possible interrupted crack propagation.

Figures 3.13 and 3.14 show the cluster locations during the 2nd and 3rd time segments, respectively. The original phone deployment plan is also plotted for comparison. The dash lines represent the hypothesized floe boundaries (Refer to Section 1.4.3). Note that the event locations, shown in Figure 3.13 and 3.14, occur at or near the hypothesized floe boundary. The events also move closer to the phones during the 3rd time segment (Figure 3.14) than those during the 2nd segment.

A probable explanation for the above observations is: the phones were deployed on Jday 164 and the floe breaks up continually with each incoming swell until the floe becomes too small to divide further. The broken pieces then collide with each other, with possible fracture at the boundary, causing the apex floe to become even smaller.

The sea state measurements confirm this finding – the swell reaches a maximum at the time period between 1800Z on J166 and 0400Z on J167 (the first 3 segments are within this time period), and then dies down. Events cluster in space at the first 3 segments and distribute more evenly at the 4th segment. These differences in the temporal and spatial distributions of events between the first three segments and the last may be due to changes in environmental forces causing the events or changes in environmental conditions, such as the swell variation and/or ice concentration fluctuation.

3.4 Event Density

The effective detection range of a hydrophone for the signal-to-noise ratio encountered is about 540 meters. The justification for this is two-fold. First, both Array #1 and Array #2 detect events located between the arrays during the 1st time segment but none during later segments. This is significant since the distance between the two arrays changes from 500m during the 1st time segment to about 1km during the next three segments. Second, the furthest event detected during the 2nd time segment by Array #2 is 533 m. The range error in event locations is about 10 meters which includes the estimation error of time delay, the travel time error in using a direct path instead of a refractive path, and the receiver position error. I therefore use 540 m as the effective detection range of a hydrophone.

Figure 3.15 shows the effective detection region for Arrays #1 and #2. The effective detection area of Array #1 is

$$A_{eff,1} = 9.7 \times 10^5 m = 0.97 km^2,$$

which includes phones D3, D5, D11, D12, and D18, and of Array #2 is

$$A_{eff,2} = 1.02 km^2$$

which includes phones D4, D8, and D19. The total detection area A_{eff} is the combination of the two:

$$A_{eff} = A_{eff,1} + A_{eff,2} - \triangle A \quad (3.1)$$

where

$\triangle A = A_{eff,1} \cap A_{eff,2}$, the common detection area of the two arrays. $\triangle A$ is $0.39 km^2$ for the first 3 segments, and is zero for the 4th Segment.

So A_{eff} is $1.60 km^2$ for the first 3 segments, and is $1.99 km^2$ for the 4th Segment.

Event density is defined as the number of events per unit area per unit time per unit frequency, *i.e.*,

$$den(f) = \frac{N(f)}{AT\Delta f} \quad (3.2)$$

where

f = peak frequency of the event,

Δf = frequency bin chosen,

T = total time of detection,

$N(f)$ = number of events with peak freq. between f and $f + \Delta f$, and

A = effective detection area of the array.

Figure 3.16 shows the event density for all time segments. Note that the event density increases with $f^{0.6 \sim 1.6}$ for $f < 300Hz$, and decreases as $f^{-1.2 \sim -2.2}$ for $f > 200Hz$. The general form for the event density is:

$$den(f) = \begin{cases} C_1 \times f^{\alpha_1}; & 10 < f < 300Hz \\ C_2 \times f^{-\alpha_2}; & 200 < f < 2000Hz \end{cases} \quad (3.3)$$

where $0.6 < \alpha_1 < 1.6$ and $1.2 < \alpha_2 < 2.2$.

The event density presented here is only a fraction of the real event density, since the latter includes weaker events whose pulse amplitude is less than the background noise. Figure 3.16 should be understood as the density for events which are detected above the background noise, i.e. this event density is a threshold dependent function. Townsend-Manning [60] has shown that more events can be detected with less background noise.

Figure 3.17 shows the ambient noise level of phone D12 at various frequency ranges (25-50, 90-100, 290-310, and 980-1000 Hz). Note that the noise levels remain constant at frequencies above 100 Hz while the noise level at 20-50 Hz increases with time. Since events are detected between 20 and 2000 Hz, the 10dB increase in 25-50 Hz bin between the 1st and 4th segment may partially account for the 10-fold event density drop between these segments.

For the first 3 segments, the peak event density is above $0.1 / km^2 / hour / Hz$ with the events clustering closely around the arrays. The 4th segment detected only 21 events, and the event clustering phenomenon diminishes.

Figure 3.1: Event Spectrum of Type I

Figure 3.2: Histogram of Peak Frequency of Events (in 100 Hz bin) within Each Time Segment

Figure 3.3: Histogram of Event Inter-Arrival Time (in 10 s bin) within Each Time Segment

Figure 3.4: Histogram of Cluster Durations (in 1 s bin) and Inter-Arrival Time (in 10 s bin) within Each Time Segment

Figure 3.5: Event Locations in the First Time Segment. The circles represent the phones of array # 1, triangles represent phones of array #2, and the diamonds represent other phones. The crosses are the event locations.

Figure 3.6: Event Locations in the Second Time Segment. The circles represent the phones of array # 1, triangles represent phones of array #2, and the diamonds represent other phones. The crosses are the event locations.

Figure 3.7: Event Locations in the Third Time Segment. The circles represent the phones of array # 1, triangles represent phones of array #2, and the diamonds represent other phones. The crosses are the event locations.

Figure 3.8: Event Locations in the Fourth Time Segment. The circles represent the phones of array # 1, triangles represent phones of array #2, and the diamonds represent other phones. The crosses are the event locations.

Figure 3.9: Phone Locations for Original Deployment during Jday 164. The floe where the phones were deployed is 400 meters in diameter, as suggested by the closed figure.

Figure 3.10: Straight Line Pattern Observed in Event Locations of Clusters in the First Segment(a). The two straight lines with arrows are marked as #1 and #2. The triangles are the receivers, and the crosses are the event locations. The broken lines are the hypothesized floe boundary.

Figure 3.11: Events Cluster in Space Observed in the First Segment.

The triangles are the receivers, and the other marks are the event locations. The broken lines are the hypothesized floe boundary.

Figure 3.12: Straight Line Pattern Observed in Event Locations of Three Clusters, which inter-arrive at swell periods, in the First Segment(b). The straight line with arrow marked as #2 is the continuation of path #2 in Figure 3.10. The triangles are the receivers, the crosses are the event locations, the broken lines are the hypothesized floe boundary, and the curly marks are the events occurred in Figures 3.10 and 3.11.

Figure 3.13: Curve Observed in Event Locations of Clusters in the Second Segment. The dashed line is the hypothesized floe boundary.

Figure 3.14: Curve Observed in Event Locations of Clusters in the Third Segment.
The dashed line is the hypothesized floe boundary.

Figure 3.15: Effective Detection Area of Arrays

Figure 3.16: Event Density for All Time Segments

Chapter 4

Figure 3.17: Ambient Noise Level Measured in Various Frequency Ranges (25-50, 90-110, 290-310, 980-1000 Hz)

Chapter 4

Source Strength: Proposed Ice Mechanisms and Acoustical Models

Since ice dominates the Arctic environment, ice mechanisms as the sources for the events are of fundamental interest and importance concerning Arctic ambient noise, for we believe that ice related events are the main contributors to the noise [16]. Nevertheless, other acoustic sources exist in the Arctic such as mammals and man-made sources. It is not difficult to distinguish these events from those which are assumed to be caused by ice. The time series of mammalian sounds are very distinctive; they sound like whines, hoots, and honks. With use of a spectrum analyzer, the frequency response of the sound made by a seal is like FM slide between 100 to 1000Hz. Man-made sounds, such as walking on the ice, are also distinctive, and often are not issues since the data analyzed are collected from drift phones which are several kilometers from the experiment team. The popping, cracking, or the combined pulse series analyzed are indeed unique and physically resembles sound generated by ice cracking, grinding, or other ice related events.

Table 4.1 lists several proposed ice mechanisms for the Arctic Ocean. They

are characterized by the ice region (the central Arctic or the Marginal Ice Zone), frequency band, event signature types, acoustic source model, and whether source motion is involved. The ridge building mechanism [53] occurs in the central Arctic, but not in the MIZ. The thermal tension crack mechanism occurs in the MIZ in seasons other than the Summer. Keenan [27] found no correlation between the ambient noise level and the air temperature above 100 Hz for MIZEX84.

Shear fracture precedes the ridge unloading motion and the wave-induced floe break up, which is demonstrated by an example shown in Figures 4.1 and 4.2. The spectrogram of the event in Figure 4.2 clearly shows that three consecutive pulses with decreasing peak frequencies, namely 345 Hz, 125 Hz, and 60 Hz. The first two pulses can be attributed to shear fractures, and the last one to the unloading motion. Shear fracture is also associated with the floe shear and collision.

The proposed ice mechanisms are based on two considerations [16]. First, the environmental forces are sufficient for the requisite deformation or motion of ice. Second, the pulse signature, radiation pattern, spectrum, and strength of derived ice events are in agreement with observations.

The following sections are devoted to acoustic models used and data analysis results for the four proposed ice mechanisms – ridge unloading, wave-induced floe breakup, floe collision and shear, and shear fracture. Dipoles and octopoles are used to represent the sources for the events. In this representation, sources should be understood as equivalent dipole forces, or equivalent octopole moments, rather than actual dipoles and octopoles.

4.1 The Ridge-Unloading Model

Dyer [16] proposed a ‘ridge-unloading’ mechanism for low frequencies based on the result that stresses due to wind, current, Coriolis force, and sea surface tilt correlated with ambient noise at 15 Hz [39]. Although the correlation was done for

Proposed Ice Mechanism	Region		Frequency Band	Event Signature Types	Acoustic Source Model	Source Motion	Discussed in Later Sections
	the CA	the MIZ					
Ridge Building	✓		mid, high	I,II,V,VI	Double Couple	Yes	No
Thermal Tension Cracking	✓	✓ except Summer	mid, high	I	Triple Couple	Yes	No
Ridge Unloading	✓	✓	low	I,II	Vertical Dipole	No	Yes (Sec. 4.1)
Wave-Induced Floe Break Up		✓	low	I,II,III	Vertical Dipole	No	Yes (Sec. 4.2)
Floe Shear and Collision		✓	extremely low	III	Double Couple	No	Yes (Sec. 4.3)
Shear Fracture	✓	✓	mid, high	I,II,V,VI	Double Couple	Yes	Yes (Sec. 4.4)

Table 4.1: Proposed ice mechanisms and their characteristics (Extremely low frequency band: < 10 Hz; low: < 100 Hz; mid: 100 to 1000 Hz; high: > 1 kHz).

pack ice, ridges exist in the MIZ. The ridge occurrence is 2.3 per km in the central Arctic, and reduces to 0.9 per km in the MIZ near the East Greenland Sea [61, Table A.3]. The ridge breaks at the accumulated moment due to the wind and current drag, and the deformed ice plate returns to its neutral buoyant position (unloading motion). This unloading motion, being an ice mechanism that involves a volume change, can be modeled by an acoustic vertical dipole.

4.1.1 Ridge Breaking

The structural response of an infinite homogeneous plate subjected to environmental moment loading on a ridge cone is first presented. Then the model discussed by Dyer [16] is used to link environmental forces with the acoustical dipole force. The ridge is modeled as a circular rigid plate of large thickness (flexural rigidity is infinite) subjected to a force couple. This circular rigid plate is attached to an infinite homogeneous thin plate, with thickness much smaller than that of the ridge.

Figure 4.3 shows the geometry for the analysis. A polar coordinate system is used; r is the radial distance from the center of the circular plate and θ is the horizontal angle from the axis of the force couple. Since the circular plate is rigid, the displacement is

$$w = w_o \frac{r}{a} \cos \theta \quad (4.1)$$

with constant slope,

$$\frac{w}{r} = \frac{w_o}{a} \cos \theta$$

where

w_o is the vertical displacement at $r = a$.

Thin plate linear elastic theory is used to solve for the response of the outer plate. The region where the radial distance is larger than a is force free. The

governing equation then corresponds to a plate on an elastic foundation [65]:

$$\nabla^4 w = -\frac{\rho_w g w}{D} \quad (4.2)$$

where

w = deflection of the infinite plate,

ρ_w = density of water on which the plate is supported,

g = gravitational acceleration,

D = flexural rigidity of the plate, $D = \frac{Eh^3}{12(1-\nu^2)}$,

E = Young's modulus of the thin plate,

h = thickness of the plate, and

ν = Poisson's ratio of the plate.

The solution of Equation 4.2 for an infinite plate is

$$w = C_1 kei(x) + C_2 ker(x) \quad (4.3)$$

where

$x = r/l$,

$kei(x)$ and $ker(x)$ = modified Bessel functions [42], and

l = flexural length scale of the plate ($= (D/\rho_w g)^{1/4}$).

The boundary conditions at $r = a$ are continuity of displacement and slope. The force equilibrium requires vanishing of total moment about the origin. Thus,

$$w|_{r=a} = w_o \cos \theta, \quad (4.4)$$

$$\frac{\partial w}{\partial r}|_{r=a} = \frac{w_o}{a} \cos \theta, \quad (4.5)$$

$$\begin{aligned} \Sigma M_o &= m - \int_0^{2\pi} (Q_r|_{r=a} + \frac{\partial M_{rt}}{\partial \theta}|_{r=a}) a^2 \cos \theta d\theta \\ &\quad - \int_0^{2\pi} M_r|_{r=a} \cos \theta a d\theta - \rho_w g \int_0^a \int_0^{2\pi} w r \cos \theta r dr d\theta \\ &= 0 \end{aligned} \quad (4.6)$$

where

m = moment about the plate center ($P \times 2a$),

Q_r = shear force,

M_r = radial moment .

Let

$$Q_r + \frac{\partial M_{rt}}{r \partial \theta} = f(r) \cos \theta$$

and

$$M_r = g(r) \cos \theta$$

then

$$\sum M_o = m - a^2 \pi f(a) - a \pi g(a) - \frac{\pi}{4} \rho_w g w_o a^3 = 0$$

The three unknowns, C_1 , C_2 and w_o , can then be determined by the three equations (Equations 4.4, 4.5, and 4.6). The results are as follows:

$$w_o = \frac{m}{I}$$

$$C_1 = A_1 w_o \cos \theta$$

$$C_2 = A_2 w_o \cos \theta$$

where

$$A_1 = \frac{\frac{l}{a} - \frac{ker^{(1)}(\frac{a}{l})}{ker(\frac{a}{l})}}{kei^{(1)}(\frac{a}{l}) - kei(\frac{a}{l}) \frac{ker^{(1)}(\frac{a}{l})}{ker(\frac{a}{l})}}$$

$$A_2 = \frac{1 - kei(\frac{a}{l}) A_1}{ker(\frac{a}{l})}$$

$$I = \frac{2\pi D}{l} II$$

and

$$II = \frac{1}{8} \left(\frac{a}{l} \right)^3 - \left(\frac{a}{l} \right)^2 \left[A_1 \left(kei^{(3)} \left(\frac{a}{l} \right) + 2 \left(\frac{l}{a} \right) kei^{(2)} \left(\frac{a}{l} \right) \right) + A_2 \left(ker^{(3)} \left(\frac{a}{l} \right) + 2 \left(\frac{l}{a} \right) ker^{(2)} \left(\frac{a}{l} \right) \right) \right]$$

The deflection w , and the radial and tangential moments, M_r and M_θ , are

$$w(x) = \frac{m}{I} \cos \theta [A_1 kei(x) + A_2 ker(x)] \quad (4.7)$$

$$M_r(x) = -\frac{m \cos \theta}{2\pi l} P_1(x) \quad (4.8)$$

$$M_\theta(x) = -\frac{m \cos \theta}{2\pi l} P_2(x) \quad (4.9)$$

where

$$\begin{aligned}
P_1(x) &= \frac{1}{II} \{ A_1 \times [kei^{(2)}(x) + \nu((\frac{l}{r})kei^{(1)}(x) - (\frac{l}{r})^2kei(x))] \\
&\quad + A_2 \times [ker^{(2)}(x) + \nu((\frac{l}{r})ker^{(1)}(x) - (\frac{l}{r})^2ker(x))] \} \\
P_2(x) &= \frac{1}{II} \{ A_1 \times [(\frac{l}{r})kei^{(1)}(x) - (\frac{l}{r})^2kei(x) + \nu kei^{(2)}(x)] \\
&\quad + A_2 \times [(\frac{l}{r})ker^{(1)}(x) - (\frac{l}{r})^2ker(x) + \nu ker^{(2)}(x)] \}
\end{aligned}$$

The radial and tangential stresses, σ_r and σ_θ , can be represented by $P_1(x)$ and $P_2(x)$ as follows:

$$\sigma_r(x) = \frac{12M_r z}{h^3} = -\frac{6mz \cos \theta}{\pi l h^3} P_1(x) \quad (4.10)$$

$$\sigma_\theta(x) = \frac{12M_\theta z}{h^3} = -\frac{6mz \cos \theta}{\pi l h^3} P_2(x) \quad (4.11)$$

where z is the distance from the central plane, $-h/2 \leq z \leq h/2$. The maximum stress occurs at the top and bottom of the ice plate. Figure 4.4 shows the ratio of the radial to tangential stress as a function of r/l for $a/l = 0.5$ and $\nu = 1/3$. Note that the ratio of σ_r to σ_θ is 3, which is $1/\nu$, at $r = a$. This is because the assumption of a rigid circular plate encircled by an infinite thin plate produces an artifact that the contact is like a clamped edge. The radial stress σ_r is larger than σ_θ for $r \leq 3a$.

Fracture occurs when the maximum tensile stress reaches the ice tensile strength, since ice is weaker under tension than under compression. The maximum tensile stress occurs at: $r = a$, $\theta = 0$, $z = -h/2$; and $r = a$, $\theta = \pi$, $z = h/2$. Material constants used are $h = 2.5$ m, $E = 9$ GPa, $\nu = 1/3$, $\rho_w g = 9000$ N/m³. The moment m is taken as 5×10^6 to 2×10^7 Pa-m (same as M_e in [16]). The deflection and the tilt angle of the rigid circular plate are 6.3 to 25.3 mm and 0.036 to 0.145 degrees respectively. From Equations 4.10 and 4.11, the maximum tensile stresses in the radial and tangential directions ($r = a$, $\theta = 0$, $z = -h/2$; and $r = a$, $\theta = \pi$, $z = h/2$) are 3.5 to 13.8 MPa and 1.2 to 4.6 MPa respectively. At $r = 3a$, the tangential stress is about equal to the radial stress, which is 0.55 to 2.2 MPa. The tensile strength of

sea ice is 0.2 to 1 MPa dependent on brine volume and temperature [64]. According to this model, the ice is likely to rupture under such forcing. The plate will suffer partial circumferential cracks at the foot and the at head of the ridge (the rigid circular plate) as the forcing reaches the ice strength.

The width of the circumferential crack can be $a \times 2\theta$, where θ is about several degrees (since the radial stress has $\cos \theta$ dependence). Before the circumferential cracks can grow wider and deeper, it will provoke radial cracks at two ends such that a wedge beam is formed. Although this fracture pattern has not been observed by the MIT field team, which was not equipped logistically or instrumentally to observe fractures directly, it is supported by some experimental observations [20,44].

Dyer [16] then considered a wedge beam attached to the tilted ridge cone. The beam deforms continuously under the moment rotating the ridge until the beam breaks at its attachment to the thin plate. The deformed beam then returns to its neutrally buoyant position, which introduces a volume change in the water. This is called the ‘ridge-unloading’ mechanism in the paper. The acoustical vertical dipole force induced by the unloading beam is

$$F = \frac{(2\pi f)^2 \rho_w l_s w_b \eta_o}{\kappa} \left[1 + \frac{1}{\kappa l_o} \right] \quad (4.12)$$

where

l_s = separation of the volume change and its image ($\approx h$),

η_o = deflection at the beam attachment to ridge cone ($= w_o$ in Equation 4.1),

$\kappa = (\rho_w g / 4D)^{1/4}$,

l_o = diameter of the ridge cone ($= 2a$),

w_b = width of the beam, and

f = characteristic frequency of the unloading motion.

Note that,

$$f \approx 2\kappa c_g / \pi \quad (4.13)$$

In his later presentation, Dyer [14] calculated the frequency dependent behavior

of the dipole force,

$$F = \left(\frac{\sigma}{E}\right) \frac{32\rho_w l_s a_2}{\pi f} \left(\frac{1-\nu^2}{2}\right) \frac{l_o^2}{h b_1} c_g^3 \left[1 + \frac{4c_g}{2\pi f l_o}\right] \quad (4.14)$$

where

σ = maximum flexural stress of the ice beam, and

a_2 = ratio of the ice beam width to the characteristic length of the ice beam.

Equation 4.14 is derived from Equation 4.12 by substituting

$$\kappa = \frac{\pi f}{2c_g}$$

$$w_b = a_2 x_e$$

$$x_e = \pi/2\kappa = \frac{\pi f}{2c_g}$$

$$\eta_o = \frac{\sigma}{2\kappa h} \left(\frac{1-\nu^2}{E}\right) \frac{l_o^2}{b_1}$$

where b_1 = parallel scale of the applied moment [16, Equation 7]. Equation 4.14 is used to compare results from the data analysis (See Section 4.1.3).

4.1.2 Equivalent Acoustic Source: Vertical Dipole Model

A vertical dipole at the sea surface is used as the source of the events (See Figure 4.5). The assumptions made in using this model are as follows:

1. Compactness: Volume change which is modeled as a monopole source occurs just below the sea surface. A compact monopole requires that the size of the source is much smaller than the wavelength, i.e. $k a \ll 1$, where k is the wavenumber and a is the radius of the source. Also the monopole source depth below sea surface is much smaller than the wavelength, i.e. $k l \sin \theta \ll 1$, where l is the separation of the source and the image and θ the launch angle at the source from the horizontal.
2. Far-field approximation: The distances between the source and the receivers are much larger than the source wavelength, i.e. $k r \gg 1$, where r is the distance from the source to the receiver.

3. Ice effect is neglected: The dipole force is applied at the free surface of the water; ice that may be present between the free surface and the water is taken to have negligible effect.

The sound pressure at the receiver due to the dipole source is

$$p(\underline{r}, t) = \frac{\sin \theta}{4\pi r} \left\{ \frac{1}{c} \frac{\partial}{\partial t} F\left(t - \frac{r}{c}\right) + \frac{F\left(t - \frac{r}{c}\right)}{r} \right\} \quad (4.15)$$

where

$p(\underline{r}, t)$ = sound pressure at $(\underline{r}, t)$,

$F\left(t - \frac{r}{c}\right)$ = dipole source function delayed by time $\frac{r}{c}$, and

c = speed of sound in water (= 1440 m/s).

For a frequency of 10 Hz, the wavelength is 144 meters. Since all phones are at a depth of 60 meters, the source range is at least 0.4 times this wavelength and $kr \geq 2.6$. Thus the second term in the equation above can be neglected throughout this thesis. Then the dipole force F is obtained by inverting Equation 4.15 for narrow band events, i.e.

$$F_{\text{peak}} = \frac{4\pi r}{k \sin \theta} p_{\text{peak}} \quad (4.16)$$

Events are treated as narrow band events since their peak frequencies are about the same as, or larger than, their bandwidths. The source strength is taken as the root-mean-square of the peak dipole force among the channels.

Two corrections have been used to obtain F_{peak} from Equation 4.16. One occurs when $kl \sin \theta$ is not small as required, and the other is caused by refractive propagation. With respect to the first, since the source depth is assumed as half the ice thickness, only sources with peak frequency larger than 1000 Hz need to be corrected, provided $\sin \theta < 0.1$ (See Appendix A.5 for more details). With respect to the second correction, refraction must be included for events beyond 300 meters of horizontal range. Figure 4.6 shows the normalized dipole field as a function of horizontal range based upon refractive propagation in a non-attenuative ocean. I

use the Safari code [54] to calculate the dipole field by placing a monopole at 1m depth, resulting in a dipole strength, $F = 8\pi$ N. The sound speed profile used for refractive propagation is based on the MIZEX84 measurement, which is shown in Figure 1.3. The normalized dipole field is:

$$\text{Normalized Dipole Field} = 10 \log \left(\frac{kF}{p(R, d)} \right)^2 [\text{dBre1m}] \quad (4.17)$$

where

k = wavenumber, $2\pi f/c$ ($c = 1440$ m/s),

f = frequency [Hz],

$F = 8\pi$ N, the dipole force,

$p(R, d)$ = acoustical pressure at (R, d) [Pa]

R = horizontal range [m],

d = receiver depth (= 60 m).

At 100 Hz, the normalized dipole field (in Figure 4.6) has a R^4 dependence below 300 m and R^2 dependence above 2000 m. The R^4 dependence is due to spherical spreading and the vertical radiation pattern, $\sin \theta$. The R^2 dependence indicates cylindrical spreading. At 1000 Hz, the normalized dipole field is 10 dB different from that of 100 Hz at 100 m and has ripples above 2000 Hz. These differences are due to $kl \sin \theta$ being not small at 1000 Hz. I use the normalized dipole field at 100 Hz to correct for refraction, and adjust for frequency by multiplying by $\sin(k(l/2) \sin \theta)/(k(l/2) \sin \theta)$ when $kl \sin \theta$ is no longer small. (See Appendix A.5)

4.1.3 Dipole Force Data

The dipole force of each event is plotted with respect to its peak frequency. Figure 4.7 shows the dipole force within each time segment. The scatter of points, over 3 to 4 orders of magnitude, prevents a satisfactory representation of dipole force by peak frequency. Figure 4.8 shows the dipole force plotted against peak frequency for all events detected in the four time segments. The result indicates a wide range as would be suggested from the previous plots. Most of the interpreted F_{peak} lies

within 3 to 4 orders of magnitude at frequencies above 100 Hz.

The spread of data in Figure 4.8 indicates the source to be more complicated than what has been considered. There may be more than one type of process responsible for the radiation at the sea surface. With the hypothesis that event signatures differentiate among the physical processes, I will use signature types to further analyze the results.

4.1.4 Signature Segregation

In the MIZ, where ice floes drift with ice concentrations between zero and unity, it is not unlikely that several different ice fracture processes are responsible for radiation of sound. The source strength of each process can have different behavior with respect to ice forcing functions, thus different peak frequencies. It is not surprising that no distinctive correlation between dipole forces and peak frequency is found (see the last section for details).

The signature type is an important characteristic of the events. It is probable that each signature type represents a different physical process, which implies that the dipole forces are directly related to the event signature types (see Figure 4.9). Note that Type I shows the least spread in the data and the dipole forces display a $f^{-0.9 \pm 0.6}$ behavior.

The dipole force of Type I events is used to compare with the central Arctic data. Dyer [16] used an empirical formula for source strength as a function of peak frequency, i.e.,

$$F_{peak} = \frac{2.5 \times 10^7}{1 + \left(\frac{f}{f_0}\right)^2} [N]; \quad f < 80 Hz \quad (4.18)$$

where $f_0 = 7.5 Hz$. The dipole forces of Type I events in the MIZ fall within the extrapolation range of the central Arctic results (See Figure 4.10). This result will be discussed in Section 4.1.6.

The relevant question is then: Is it possible that dipole forces depend upon

parameters other than the peak frequency? To discover the possibility, I use Type II & III events for further analysis in the next section.

4.1.5 Dipole Force vs. Decay Rate of Type II & III

Signature of Type II and III events resemble damped sinusoids and can be represented mathematically as follows:

$$p(t) = A \exp(-\alpha t) \sin(2\pi f_o t), \quad t > 0 \quad (4.19)$$

where

A = magnitude of the peak,

f_o = peak frequency [Hz], and

α = decay rate [s^{-1}].

The decay rate α can be estimated by measuring the envelope, i.e.,

$$\alpha = f_o \ln \frac{p_1}{p_2} \quad (4.20)$$

where p_i is the oscillation magnitudes of the form in Equation 4.19. Equation 4.20 can be extended to n successive magnitudes.

Figure 4.11 shows the results for Type II and III events during the 3rd and 4th segments. The scatter is about two orders of magnitude, from 1 to 100 kN for $10 < \alpha < 2000 \text{ s}^{-1}$. It seems that no clear relation exists between the dipole force for ‘damped sinusoid’ events and their decay rate.

4.1.6 Discussion on Dyer’s Modified Ridge-Unloading Model

This section presents the fitting of data to the ice model of Section 4.1.1. Dyer used Type I events in Figure 4.9 and results from Townsend-Manning [60], which are mostly Type I events, to fit the ridge-unloading model [14]. The dipole force F of Equation 4.14 is plotted vs. f in Figure 4.12 with the parameters as follows:

$$\begin{aligned} h &= 2.5\text{m}, & l_s &= 2.5\text{m}, & a_2 &= 0.1(2/\pi), \\ \sigma/E &= 3 \times 10^{-4}, & l_o &= 25\text{m}, & b_1 &= 25\text{m} \end{aligned}$$

The figure also shows the dipole forces of Equation 4.18 as the dash line, which agrees with dipole forces of Type I events from both the MIZ (Figure 4.12) and the central Arctic. The curve of Equation 4.14 is 10 times more at 10 Hz and 1000 times more at 1000 Hz than the data. Also the frequency dependence of Equation 4.14 is not in agreement with the data. Dyer then modified the model by including a creep effect such that the ratio of the ice strength under the creep effect, to the ice strength of linear elasticity has frequency dependence of $\sigma_{\text{creep}}/\sigma_{\text{elastic}} \sim f^{-2}$. This modification gives good agreement for frequencies below 200 Hz. The fact that the data scatter at frequencies above 100 Hz and the poor fit of data with theory above 200 Hz indicate that there may be other mechanisms involved for frequencies above 100 Hz.

4.2 Wave-Induced Floe Breakup Model

Wave-induced floe breakup is a very likely source of ambient noise in the MIZ. Wadhams [61] concluded that breakup of ice floes due to flexure caused by surface gravity wave motion is the major factor in determining the maximum size of an ice floe that can exist in the MIZ. The breakup occurs near the ice edge, since waves attenuate very highly once they enter the ice covered zone.

The breakup may be the result of several cycles of wave-induced flexure, a fatigue failure. The first stage involves stable cracks propagating intermittently. Then, as the fracture toughness of the floe reduces, with increases in the crack size, the crack propagation becomes unstable and the crack propagates through the depth and length of the floe. After the crack breaks through, the broken floes (smaller in size) adjust themselves to the wave condition. Both the crack propagation and the position change before and after the breakup cause acoustic radiation into the water.

An ice floe breakup due to incoming wave was modeled by Goodman [61, Section B8]. He took this as a two-dimensional problem by assuming an infinitely long strip

Half of Crack Length, (l_c)	Critical Strain (ϵ_{crit}),
1 mm(grain size)	1.9×10^{-5}
10 cm	1.9×10^{-6}
1 m	5.4×10^{-6}

Table 4.2: Critical strain for various crack sizes

of finite width (I will call the finite width of the strip as floe length in the following text) with waves incoming from the direction of the finite width. The results are plotted in Figure 4.13 as the vertical strain at mid floe normalized by the surface wave amplitude with respect to the floe length, i.e., $|\epsilon|/A'|_{x=l/2}$ versus l , for three wave periods (12 s, 15 s, and 17 s). The amplitude of the ice coupled wave, A' , is about 0.9 times the wave amplitude [61]. The fracture criterion for an ice floe with a pre-existing crack of $2l_c$ long is that the strain exceed ϵ_{crit} ,

$$\epsilon_{crit} = \frac{K_{IC}}{\beta\sqrt{\pi l_c}} \frac{(1 - \nu^2)}{E} \quad (4.21)$$

where

K_{IC} = fracture toughness of ice (100 kN- $m^{-3/2}$ [43]),

β = constant, of order of 1,

ν = Poisson's ratio, and

E = Young's modulus.

Using $E = 9$ GPa, $\nu = 1/3$, and $\beta = 1$, the critical strain for various crack sizes is shown in Table 4.2.

Based on Table 4.2 and Figure 4.13, the minimum wave amplitude required to cause a pre-existing crack to propagate in an ice floe can be calculated as a function of floe length. For the same floe length and at the same wave period, the minimum amplitude of the ice-coupled wave is

$$A'_{min} = \frac{\epsilon_{crit}}{(\epsilon_{mid-floe}/A')}$$

Table 4.3 shows that for a pre-existing crack 20 cm long and a wave period of

Floe Length			400 m	200 m	100 m
Wave Period [s]	Half of Crack Length, l_c	ϵ_{crit}	A'_{min} [m]	A'_{min} [m]	A'_{min} [m]
12	1 mm	1.9×10^{-5}	0.0400	0.0380	0.189
12	10 cm	1.9×10^{-6}	0.004	0.0038	0.019
12	1 m	5.4×10^{-6}	<u>0.001</u>	<u>0.001</u>	<u>0.006</u>
15	1 mm	1.9×10^{-5}	0.054	0.069	0.4200
15	10 cm	1.9×10^{-6}	0.005	0.007	0.042
15	1 m	5.4×10^{-6}	0.002	0.002	0.013
17	1 mm	1.9×10^{-5}	0.076	0.108	0.756
17	10 cm	1.9×10^{-6}	0.007	0.011	0.076
17	1 m	5.4×10^{-6}	0.002	0.003	0.025

Table 4.3: The minimum surface wave magnitude needed to cause additional fracture in an ice floe of various length (400 m, 200 m, and 100 m) with various crack sizes (1 mm, 10 cm, and 1 m) at various wave periods (12 s, 15 s, and 17 s). The three underlined values are the smallest A'_{min} for the three different floe lengths.

12 s, the smallest surface wave needed to cause fracture in an ice floe of 400 m long is only 4 mm in amplitude, 3.8 mm for a 200 m floe, and 19 mm for a 100 m floe.

There is no measurement of swell period and magnitude corresponding to the time and locations of the data analyzed. So I use an indirect measurement of swell magnitude by Martin et al. [40] as an indicator of the possibility of ice fracture due to incoming swell. The swell root-mean-squared amplitude measured by Martin et al. varied between 1 mm and 30 mm during June 8 to June 16. Their measurements are for floes 5 km from the Polar Queen (a research ship of MIZEX84), which was about 42 km away from Kvitbjorn (the acoustic array is less than 7 km from Kvitbjorn). Martin et al. [40] also observed that the swell periods are between 10 s and 18 s. These measurements show that the swell is strong enough to cause fracture in an ice floe. The fracture process will be discussed further in Section 4.4.

I calculate the floe displacement before and after breakup for different floe lengths (100 m, 200 m, and 400 m) with an incident wave period of 12 s. The breakup position is at the half-length point where the maximum strain occurs. Figure 4.14 shows the floe displacement before and after breakup. The after-position is calculated by assuming a floe of half size under the same wave condition. The characteristic frequency is defined as the frequency of the flexural wave traveling from the original floe to the new floe of half size, i.e.,

$$f = c_g / (l/2) \quad (4.22)$$

where l is the original floe size and c_g is the group speed of the flexural wave of the ice (refer to Figure 2.24 of [59]). For a 3 m thick ice floe, c_g is approximately

$$c_g = 200f^{0.6} \text{ m/s}; \quad f < 30 \text{ Hz},$$

$$c_g = 1414 \text{ m/s}; \quad f = 100 \text{ Hz},$$

$$c_g = 1200 \text{ m/s}; \quad f > 300 \text{ Hz}.$$

The floe length versus the peak frequency for an unloading motion when an ice floe breaks in the middle is shown in Figure 4.15. For frequencies above 100 Hz, the

floe length is less than 30 m. The swell has to be 0.6 m high to break an ice floe of 30 m long with a 2 m pre-existing crack. For frequencies below 30 Hz, the floe length is larger than 100 m, which is likely to have fracture due to incoming swell. Thus I conclude that the unloading motion of ice floe breaking due to swell should be one of the causes for acoustical events below 30 Hz.

The volume changes involved in the unloading motion of floe breakup can be modeled as several vertical dipoles along the floe length at the horizontal plane. The arrows in Figure 4.14 represent the directions and positions of the vertical dipoles.

4.3 Floe Shear and Collision Model

Floe shear and collision, or floe-floe interaction, are mechanisms in the marginal ice zone, which occur during ice motions. The floe differential motions are due to environmental forces such as waves (surface gravity or internal wave), wind, current, Coriolis force, and tides. Ice converges and diverges in response to these forces. Martin et al. reported that the floe collision energy correlates with the swell energy ([40], or in Section 3.3). Reports on mechanisms of ice/ocean eddies generation ([63,23,56]) suggest that baroclinic and barotropic instabilities and topographic inhomogeneities have effects on ice motions. Internal waves are suggested by Muench et al. [50] to be the cause of ice banding. Interactions of the internal waves with the ice are confirmed by correlation of the ambient noise at low frequencies (25 to 50 Hz) with internal waves as reported by Makris [38], which also indicates that floe interactions caused by internal waves during ice motions are possible mechanisms for noise events.

This section is divided into two parts: floe collision (or normal motion) and floe shear (or tangential motion).

4.3.1 Floe Collision

The process of drifting ice floes colliding with bridge piers, or other structures, has been reported by Michel [44, Section 4.2] and others [57]. Michel describes the process of a drifting ice floe colliding with a bridge pier in ductile or brittle response depending on the effective strain rate, floe velocity divided by 4 times the width of the contact area ($\dot{\epsilon} = V/(4b)$). For $10^{-8} < \dot{\epsilon} < 5 \times 10^{-4} s^{-1}$, the ice has ductile behavior when impacting a rigid structure and micro cracks are formed within the damage zone (within a range of $2.5b$, where b is the width of the contact area). In brittle response ($\dot{\epsilon} > 10^{-2} s^{-1}$), major cracks are formed and ice will cleave along the cracks. The effective ice force exerted on the structure is

$$F_{crush} = \sigma_{crush}bh \quad (4.23)$$

where

σ_{crush} = ice crushing strength,

b = width of the contact area, and

h = ice floe thickness.

F_{crush} has been studied in laboratory ice/structure interaction. Sodhi [57] measured the ice force on the rigid structure by moving a rigid cylinder against an ice sheet at constant speed. The force has a cyclic behavior which depends on strain rate. Each cycle represents a complete crushing of a damage zone.

In the marginal ice zone, ice floe collisions occur frequently. Two ice floes slowly approach each other and after initial contact, the edges crush and pieces of ice are broken. This process is similar to ice floes colliding with a bridge pier with brittle failure, i.e. for $\dot{\epsilon} > 10^{-2} s^{-1}$. During the collision process, one or both floes suffer cracking and breaking, and both are subject to the force which is the contact area multiplied by the ice crush strength (Equation 4.23). The collision force also equals the ice mass multiplied by the collision acceleration. Using the acceleration measurement by Martin et al. [40], which is of the order of 5 mm/s^2 , the collision

force of two ice floes of equal size and acceleration is

$$F_{collision} = 2\rho_i L^2 h a \cong 12.25 L^2 [N] \quad (4.24)$$

where

ρ_i = ice density (980 kg/m³),

L = ice floe length and width [m],

h = ice thickness (2.5 m), and

a = collision acceleration ($5 \times 10^{-3} m/s^2$).

Thus $F_{collision}$ is 2450 N for $L = 10$ m, and is 2.45×10^5 N for $L = 100$ m.

Assuming the ice contact extends the whole thickness, the maximum width of the contact area when the ice crushes under $F_{collision}$ is

$$b_{maximum} = \frac{F_{collision}}{\sigma_{crush} h} = 3.0 \times 10^{-6} L^2$$

where $\sigma_{crush} = 1.5$ MPa, ice crushing strength at a brittle range and at 0°C [44].

Thus b is 0.3 mm for a 10 m wide floe and 3 cm for a 100 m wide floe.

Using the collision velocity measured by Martin et al. [40], about 1 mm/s, the effective strain rate of a collision process is

$$\dot{\epsilon} = \frac{2.5 \times 10^{-4}}{b} = \frac{83}{L^2}$$

Thus $\dot{\epsilon} = 0.83 s^{-1}$ for a 10 m wide floe and $8.3 \times 10^{-3} s^{-1}$ for a 100 m wide floe.

Hence ice collisions lie beyond the ductile region and are within the transition to brittle regions. The size of the fractured ice is [44, Section 4.2.2.2]

$$0.25 < \delta < 0.5h$$

The frequency of the fracture is then

$$f = \frac{v}{\delta}$$

where v = the propagation speed of the fracture. Broek [4] derived an estimate of the upper boundary of crack speed as $0.38 c_{ice}$, where c_{ice} is the sound speed of the

material. This upper-bound of crack speed was also supported by measurements for brittle cracks. The sound speed of ice is about 3200 m/s and the Rayleigh wave speed of ice is 1700 m/s. Based on Broek's derivation, The fracture propagation speed of ice is assumed to be equal to or less than $0.7c_R$, where c_R is the Rayleigh wave speed of ice. Then f is in the range of several hundred Hz and above, i.e. $f > 800$ Hz. Radiation from a fracture due to floe collision is discussed in Section 4.4.

Buckingham and Chen [7] have proposed a collision model which assumed the ice floe as a liquid sphere surrounded by water, and replaced the crush area by putting a monopole in its center. The full fluid wave equations were then solved. Several issues deserve further attention:

1. The period between successive pulses is modeled as $2l/c_\alpha$, where l is the floe length, and c_α is the longitudinal wave speed of ice. With use of $c_\alpha = 3000\text{m/s}$ and $l = 100\text{m}$, the period is 0.06 s, which is larger than the pulse separation of Type V & VI events.
2. The decay ratio of the successive pulses should include attenuation which is 0.5 dB/m at 300 Hz [59, Figure 2-15]. Then the decay ratio is 60 dB for a 100 m long floe. Successive pulses basically could not exist with such high longitudinal wave attenuation.
3. The fracture is modeled better as a double couple than a monopole (see Section 4.4).
4. We assume that all pulse radiations are the same to predict the ambient noise level, which contradicts observations that events vary in their peak frequencies and signature types.

I think the shear fracture model (See Section 4.4) is more appropriate in modeling the fracture due to floe collisions.

4.3.2 Floe Shear

Two processes are discussed in this section: the fracture of asperities and the normal vibration of asperities when floes shear against each other. The edges of ice floes are characterized by asperities. The asperity on the edge of an ice floe during a floe shear can be modeled as a cantilever beam mounted on the ice with a point force applied at the free end. The force is the tangential frictional force μF_n , where μ is the frictional coefficient (about 0.1), and F_n is the normal force, $F_n = F_{collision}(D/d)^2$, where $F_{collision}$ in Equation 4.24 is assumed to be the total normal load, D is the asperity spacing, and d is the asperity width. The maximum stress occurs at the fixed end,

$$\sigma = \frac{Pl d/2}{I} \left(\frac{D}{d}\right)^2$$

where

P = tangential force ($= \mu F_{collision}$),

l = length of the asperity,

d = width of the asperity (assume the asperity has a square cross section),

$I = d^4/12$.

Substituting the above into Equation 4.24, the maximum stress becomes

$$\sigma = 73.5 L^2 \frac{l}{d^3} \left(\frac{D}{d}\right)^2$$

The fracture occurs at the fixed end when σ reaches the flexural ice strength (about 0.2 to 1 MPa). Therefore,

$$\frac{l}{d^3}_{max} = \frac{\sigma_{crit}}{73.5 L^2 \left(\frac{D}{d}\right)^2}$$

Assume D/d is of the order of 10, then for $L = 100$ m, l/d^3 is about 0.01, hence, $d = 0.5$ m at $l = 1$ mm, $d = 1$ m at $l = 1$ cm, and $d = 2$ m at $l = 10$ cm. These dimensions are within a reasonable physical range. The characteristic frequency of the fracture at the fixed end is

$$f = \frac{v}{d}$$

where v is the fracture propagation speed ($\leq 0.7c_R$) [4]. The range of d is from 0.5 m to 2 meters, so f is in the range of several hundred Hz and above. The radiation model for this fracture process is discussed in Section 4.4.

The normal vibration of asperities during floe shears is similar to the normal frictional vibrations of bearings described by Massouros et al. [41]. I use their model to do a rough estimation of the characteristic frequency, which is [41, Equation 28]

$$f_n = \frac{1}{2\pi} \sqrt{c_{12}/m} = \frac{1}{2\pi} \sqrt{\frac{W}{m} \frac{Z_1 Z_2}{Z_1 h_{o2} + Z_2 h_{o1}}}$$

where

c_{12} = resultant contact rigidity coefficient,

m = vibrating mass ($m_1 + m_2$),

W = normal load,

Z_i = a function of the ratio of the real contact area to the apparent contact area ($Z_i \leq 6$), and

h_{oi} = height of the asperities.

Assume the normal load $W = (m_1 + m_2)a$, where a is the collision acceleration (about 5 mm/s^2 [40]), and

$$\frac{Z_1 Z_2}{Z_1 h_{o2} + Z_2 h_{o1}} \sim \frac{Z}{h_o}$$

Then

$$f_n = \frac{1}{2\pi} \sqrt{\frac{a}{h_o} Z}$$

where

h_o is of the same order as h_{oi} , and Z is the smaller of Z_1 and Z_2 . For $h_o = O(mm)$ and $Z = 6$ (the maximum value), f_n is about 2 Hz. The normal vibration due to floe shear cannot be a source mechanism for events above 10 Hz.

4.4 Shear Fracture Model

The MIZ is composed of broken floes of various sizes. Winds, currents, and waves drive the floes to converge, or diverge, causing them to shear and bump against

each other which results in fracture of ice. This can be modeled as force couples acting on the fault plane of an ice sheet, similar to earthquake seismology. Shear fracture, a part of wave induced fracture (Sec. 4.2), is precursive to ridge unloading motion (Sec. 4.1). This shear fracture model presented in this section takes all these mechanisms into account.

The basic three modes in fracture – opening, tearing, and slipping – result from different force couple arrangements. Figure 4.16 shows the nine possible force couples or quadrupoles. The shear fracture model, discussed in this section, is mainly the tearing and slipping modes. Similar to the simplified dipole model, force couples acting on the fault plane can be modeled acoustically as quadrupole moments transforming into octopoles due to the free surface image. Dyer [14] uses a compact octopole to model tangential motion in the vertical direction of ice floes under shear. The octopole source of a narrow band event is

$$S_o = k^3 M_o l_M \quad (4.25)$$

where

k = wavenumber (w/c),

l_M = distance between the moment and its image ($l_M \sim h$),

h = ice thickness, and

M_o = double couple moment (or fracture moment).

In seismology, M_o is also called the seismic moment [5,2],

$$M_o = \mu A \bar{u} \quad (4.26)$$

where

μ = shear modulus of the ice,

A = area of the fault plane, and

$\bar{u}$ = average displacement over fault plane.

With $\lambda = \mu$ and $A = \pi r^2$, the average dislocation becomes two-thirds of the maxi-

mum final dislocation for 100% stress drop. That is,

$$\bar{u} = \frac{16}{7\pi} \frac{\sigma}{\mu} r$$

The corresponding fracture moment is

$$M_o = \frac{16}{7} \sigma_B \left(\frac{A}{\pi} \right)^{\frac{3}{2}} \quad (4.27)$$

where

r = radius of the fault plane, and

σ_B = ‘Brune Stress Drop’.

The following section presents an analysis of the shear fracture for different orientations, which can also be represented by Equations 4.25, 4.26, and 4.27. In the model, octopoles are used to analyze the source strengths of the events.

4.4.1 Octopole Model and Source Strength

The octopole model is essentially a double couple transformed into an octopole by the free surface (a double couple corresponds to force couples in seismology [2]). The assumptions in the octopole model are:

1. Compactness: There is a double couple applied below the free surface, which becomes an octopole due to the free surface. The separation between the double couple and its image is much smaller than the wavelength of the source, more precisely $kl_M \sin \theta \ll 1$.
2. Far Field Approximation: The receiver range is much larger than the wavelength of the source, $kr \gg 1$.
3. No Ice: The double couple is applied immediately below the free surface and acts on the water, i.e. the existence of the ice is neglected.

Figure 4.17 shows the spherical coordinate system used. The positive z-direction points downward, the x-axis is either parallel or vertical to the force separation, the

angle θ is measured from the x-y plane, the horizontal plane, and ϕ is the angle from the x-axis.

The shear fracture is modeled by a double couple, (i, j) and (j, i) , to represent a fault plane normal in the j -direction moving along the i -direction, or the plane normal in the i -direction moving along the j -direction. Three double couples (See Figure 4.16) are used here – $(1,2)$ & $(2,1)$, $(1,3)$ & $(3,1)$, and $(2,3)$ & $(3,2)$. The octopole radiation due to a double couple [28] is:

$$p(r, t) = \frac{g(\theta)B(\phi)}{2\pi r} \left\{ \frac{1}{c^3} \frac{\partial^3}{\partial t^3}(Mh) + \frac{3}{rc^2} \frac{\partial^2}{\partial t^2}(Mh) + \frac{6}{r^2c} \frac{\partial}{\partial t}(Mh) + \frac{6}{r^3}(Mh) \right\} \quad (4.28)$$

Only the first term is needed under the far-field approximation, thus

$$p(r, t) = \frac{g(\theta)B(\phi)}{2\pi r} \frac{1}{c^3} \frac{\partial^3}{\partial t^3}(Mh) \quad (4.29)$$

where

$g(\theta)$ = vertical directivity,

$B(\phi)$ = horizontal directivity,

c = sound speed in the water,

r = slant range,

h = twice the source depth below the sea surface (= ice floe thickness), and

M = force couple moment ($= F \times l_M$).

I then divide the three possible octopoles into two groups based upon the vertical directivity [28],

$$\text{Group I (lat1): } g(\theta) = \sin \theta \cos^2 \theta$$

$$\text{Group II (lat2): } g(\theta) = \sin^2 \theta \cos \theta$$

These are so-called lateral octopoles which include three double couples (See Figures 4.18 and 4.19 for the geometry). The octopole of Group I is transformed from the double couples, $(1,2)$ & $(2,1)$ (See Figure 4.18 and Figure 4.16). The two double couples shown in Figure 4.19 ($(1,3)$ & $(3,1)$ and $(2,3)$ & $(3,2)$) are related to the two octopoles of Group II. In group I, the horizontal directivity pattern is

$$B_{lat1,1}(\phi) = \sin(\phi) \cos(\phi)$$

Group	$g(\theta)$	$B(\phi)$	Moment (M)	Double Couple
I	$\sin \theta \cos^2 \theta$	$\sin(\phi) \cos(\phi)$	$F \times l_{1,2}$	(1,2) & (2,1)
II	$\sin^2 \theta \cos \theta$	$\sin \phi$	$F \times l_{2,3}$	(2,3) & (3,2)
II	$\sin^2 \theta \cos \theta$	$\cos \phi$	$F \times l_{1,3}$	(1,3) & (3,1)

Table 4.4: Summary of characteristics of octopoles

In group II, two possible values for $B(\phi)$ exist. For the double couple (2,3) & (3,2),

$$B_{lat2,1}(\phi) = \sin(\phi)$$

For (1,3) & (3,1),

$$B_{lat2,2}(\phi) = \cos(\phi)$$

Table 4.4 summarizes the above.

For a narrow band event, the pressure can be approximated as

$$p_{peak} = \frac{g(\theta)}{2\pi r} B(\phi) k^3 M h \quad (4.30)$$

The octopoles require one extra source parameter, namely the source orientation. For measurements, we need sensors distributed in the horizontal plane about the source to estimate the source orientation. It is almost impossible to obtain such a data set in a MIZ field experiment where natural sources are to be studied with hydrophones whose drift typically concentrates them. In practice, the sensors are usually distributed in one sector, or often bounded to about a 90 degree angular sector. Without knowledge of the orientation of the source, I therefore include $B(\phi)$ in the source term and evaluate the octopole source strength as

$$\begin{aligned} S_{octo} &= B(\phi) k^3 M h \\ &= p_{peak} \frac{2\pi r}{g(\theta)} [N/m] \end{aligned} \quad (4.31)$$

Correction for refraction is done for all data by using the normalized dipole field curve, at 100 Hz (Figure 4.6) divided by $g(\theta)/\sin \theta$. Correction for high frequencies where $kl_M \sin \theta$ is not small is included in all the data as in the dipole force calculation (Appendix A.5 includes the details).

The root-mean-squared values of S_{octo} for all channels are used on all plots. The variation in S_{octo} is mainly due to $B(\phi)$. The absolute value of $B(\phi)$ can vary by several orders of magnitude.

Figure 4.20 shows the octopole source strength of Group I octopoles vs. peak frequency for all events of Type I. (There are 25 Type I events in all four time segments.) Values of $S_{octo,lat1}$ lie between 10^3 to 10^5 N/m for all Type I events. A least-square-fit straight line of the log-log data points has a positive slope of 0.50 with an error of 0.47.

$$S_{octo,lat1,typeI} = 10^{2.7} \times f^{0.50 \pm 0.47} [N/m] \quad (4.32)$$

Figure 4.21 shows the octopole source strength of Group II, $S_{octo,lat2}$ for all Type I events, vs. their peak frequencies. The spread of $S_{octo,lat2}$ is due to S_{octo} of events located more than 100 m away. These events have larger $S_{octo,lat2}$ than $S_{octo,lat1}$, due to the dependence on θ . As shown previously, $g(\theta)$ is a function of $\sin \theta$ and $\cos \theta$. Since $\theta < 45^\circ$ for all events (the event are located at least 60 m away in the horizontal plane), $\cos \theta$ is $1/\sqrt{2} \sim 1$. Thus the differences in S_{octo} for events far away are caused by the $\sin \theta$ dependence. The relationships of S_{octo} for the two octopole groups are then, approximately

$$\frac{S_{octo,lat1}}{\sin \theta} = \frac{S_{octo,lat2}}{\sin^2 \theta} \quad (4.33)$$

For events of range smaller than 100 m, $\sin \theta$ and $\sin^2 \theta$ are within the same order (0.1 to 1). On the other hand, for events of range of 600 m, $\sin \theta$ is 10 times larger than $\sin^2 \theta$. In other words, $S_{octo,lat1}$ is then 10 times smaller than $S_{octo,lat2}$ for events located 600 m away.

Figures 4.22 and 4.23 present the S_{octo} of Type II and III events of the two octopole groups. Type II and III events have S_{octo} between 1 and 1000 [kN/m] for most of the events. The differences between S_{octo} of Type I events and Type II and III events are:

1. $S_{octo,lat1}$ exhibits a $f^{1.4 \pm 0.24}$ frequency dependence for Type II & III events (See Figure 4.22) instead of $f^{0.50 \pm 0.47}$ for Type I events, i.e.

$$S_{octo,lat1,type2-3} = 10^{0.46} \times f^{1.4 \pm 0.24} \quad (4.34)$$

2. S_{octo} of all Type II & III events lies between 1 and 1000 kN/m instead of 1 to 100 kN/m for Type I events.

4.4.2 Effect of Neglect of Ice in the Octopole Model

The use of an octopole as the source model in this thesis excludes the effect of ice. Figure 4.24 compares the octopole model with Kim's result [29] which includes the ice and refraction with use of a mild SSP for the CA. The curves of my octopole model has no refraction, however, refraction is not important at horizontal range 300 m with use of the sound speed profile in Figure 1.3. The curves named 'strike-slip' and 'dip-slip' are taken from Figure 6.9 of [29]. Kim placed double couples in a continuum model to determine the acoustical radiation from ice motion at a pre-existing fault plane. The inputs for these curves are as follows: the normal of the fault plane is in the horizontal direction, the receiver position is $(R, \theta, z) = (300 \text{ m}, 30^\circ, 60 \text{ m})$ of a cylindrical coordinate system. The transfer function is defined as

$$S_p(f) = |H(f)|^2 \times \left(\frac{|M(f)|^2}{4\pi} \right) \quad (4.35)$$

where $S_p(f)$ is the spectrum of the acoustic radiation received at the hydrophone, $M(f)$ is the seismic moment, the source term. The fault plane is in the x-z plane and the principal axis of the plane is in the x-direction. The motion of the fault plane for the strike-slip curve is in the x-direction (parallel to the principal axis of the fault plane), and for dip-slip is in the z-direction (normal to the principal axis of the fault plane). The transfer function of the octopoles derived from Equation 4.30 is

$$H(f) = \frac{2g(\theta)B(\phi)k^3h}{r} \quad (4.36)$$

In Figure 4.24, the curve marked as ‘Octopole I’ is the octopole corresponding to strike-slip. The difference between the two is less than 10 dB in the frequency range of 10 to 200 Hz. The transfer function of the strike-slip curve has f^3 dependence in this range which is same as the octopole. Then the transfer function decreases beyond 150 Hz until it starts to level off at frequencies above 300 Hz. So the Octopole I curve goes beyond the strike-slip curve by 20 dB for frequencies larger than 300 Hz. The dip-slip and ‘Octopole II’ curves shown have a different behavior than the previous. The difference between the Octopole II and dip-slip curves never exceeds 15 dB for the frequency range of 10-300 Hz. The dip-slip curve increases as f^4 in this range. It starts to level off at frequency larger than 300 Hz. Then the difference is 18 dB at 1000 Hz and 32 dB at 2000 Hz.

The ice makes the acoustic radiation stronger for the strike-slip motion by 10 dB for frequencies between 10 and 200 Hz, and weakens the acoustic radiation by more than 20 dB above 300 Hz. That is, if I use the Octopole I model for source inversion, I overestimate $S_{octo,lat1}$ 3-fold at frequencies between 10 and 200 Hz, and underestimate it 10-fold at frequencies beyond 300 Hz. For the dip-slip motion, the ice effect strengthens the acoustic radiation by less than 10 dB between 30 and 500 Hz, weakens it by less than 10 dB between 10 and 30 Hz, and weakens the radiation by $60 \log(f/500)$ dB beyond 500 Hz. Thus I overestimate $S_{octo,lat2}$ less than 3-fold between 30 and 500 Hz, and underestimate it by $(f/500)^3$ times at frequency beyond 500 Hz. This comparison indicates that the ice cannot be neglected at frequencies above 300 Hz; the wavelength is comparable with the ice thickness at these higher frequencies.

4.4.3 Propagating Rupture

The octopole model assumes the source is motionless and compact. For propagating cracks, the zero motion assumption may no longer be valid. The propagating rupture can be modeled as a phase-tapered array [2,29].

The propagating rupture is modeled as follows: Assume a fault plane of width W and length L , the rupture starts at one end and propagates to the other at speed v , essentially a line array of compact octopoles with time delays ξ/v , where ξ is the position of an octopole measured from the starting point of rupture. The far field radiation is the integration of Equation 4.29 over L with uniformly distributed moment, $m = M/L$, such that

$$m = \begin{cases} \frac{M(t - \frac{\xi}{v})}{L}, & -L/2 \leq \xi \leq L/2 \\ 0, & \text{otherwise} \end{cases}$$

Thus

$$p(\underline{r}, t) = \int_{-L/2}^{L/2} \frac{g(\theta)B(\phi)}{2\pi r} \frac{1}{c^3} \frac{\partial^3}{\partial t^3} \left[\frac{M(t - \frac{r_o}{c} + \frac{r'}{c} - \frac{\xi}{v})h}{L} \right] d\xi \quad (4.37)$$

where $r = r_o - \xi \cos \theta \cos(\phi - \phi_d)$,

$\frac{r_o}{c}$ = time delay due to the distance from $\xi = 0$ to the receiver,

ϕ_d = horizontal angle of a line array

$\frac{\xi}{v}$ = time delay due to the propagation at speed v .

$\frac{r'}{c}$ = time delay due to a signal from the difference between $\xi = 0$ and ξ to the receiver, i.e

$$\frac{r'}{c} = \frac{\xi \cos \theta \cos(\phi - \phi_d)}{c}$$

Consider

$$\alpha = \frac{1}{v} - \frac{\cos \theta \cos(\phi - \phi_d)}{c}$$

$$1/r \sim 1/r_o$$

$$t' = t - \frac{r_o}{c}$$

and

$$S(t' - \alpha\xi) = \frac{1}{c^3} \frac{\partial^3 M(t' - \alpha\xi)}{\partial t^3}$$

Then Equation 4.37 becomes

$$p(\underline{r}, t) = \frac{g(\theta)B(\phi)h}{2\pi r L} \int_{-L/2}^{L/2} S(t' - \alpha\xi) d\xi \quad (4.38)$$

With $\tau = \alpha\xi$ ($d\tau = \alpha d\xi$), Equation 4.38 becomes

$$\begin{aligned} p(\underline{r}, t) &= \frac{g(\theta)B(\phi)h}{2\pi r L} \int_{-L\alpha/2}^{L\alpha/2} S(t' - \tau) d\left(\frac{\tau}{\alpha}\right) \\ &= \frac{g(\theta)B(\phi)h}{2\pi r L} \int_{-\infty}^{\infty} S(t' - \tau) \times w(\tau) d\tau \\ &= \frac{g(\theta)B(\phi)h}{2\pi r L} S(t') * w(t') \end{aligned}$$

where

$$w(\tau) = \begin{cases} 1/\alpha, & -L\alpha/2 \leq \tau \leq L\alpha/2 \\ 0, & \text{otherwise.} \end{cases}$$

and the symbol $*$ denotes linear convolution of two functions. Convolution of two functions in the time domain is equivalent to multiplying the Fourier transforms in the frequency domain, i.e.

$$p(\underline{r}, f) = \frac{g(\theta)B(\phi)h}{2\pi r L} S'(f)W(f) \quad (4.39)$$

where

$$\begin{aligned} S'(f) &= \mathcal{F}\{S(t)\} \exp(j2\pi f r_o/c) \\ W(f) &= L \frac{\sin(2\pi f L\alpha/2)}{2\pi f L\alpha/2} = L \frac{\sin X}{X} \\ X &= \pi f \alpha L \end{aligned}$$

The spectrum of $p(\underline{r}, t)$ is then

$$S_p(\underline{r}, f) = \left[\frac{g(\theta)B(\phi)h}{2\pi r} \right]^2 \left(\frac{\sin X}{X} \right)^2 |S(f)|^2$$

For a narrow band event,

$$S(f) = \mathcal{F}\left\{ \frac{1}{c^3} \frac{\partial^3 M(t)}{\partial t^3} \right\} = k^3 M(f)$$

Then

$$S_p(\underline{r}, f) = \left[\frac{g(\theta)B(\phi)k^3 h}{2\pi r} \right]^2 \left(\frac{\sin X}{X} \right)^2 |S_M(f)|^2 \quad (4.40)$$

where $S_M(f)$ is the spectrum of $M(t)$. The nodes of $|\frac{\sin X}{X}|^2$ are located at $f = n/(\alpha L)$, with n being an integer. The null spacing is:

Phone#	ϕ	f_n (Hz)	f_{n+1} (Hz)	Δf	$\cos \Phi$	Φ
D3	340°	380	600	220	-0.27	-106°
D5	59°	380	620	240	-0.17	100°
D11	56°	380	600	220	-0.27	106°
D12	46°	850	1030	180	-0.56	124°
D18	61°	620	880	260	-0.08	95°

Table 4.5: Calculation of null spacings and the source orientation

$$\Delta f_{null} = (\alpha L)^{-1}$$

Assume the peak frequency of a event to be

$$f_p = \frac{v}{L}$$

then

$$\Delta f_{null} = \frac{f_p}{1 - (v/c) \cos \theta \cos \Phi}$$

where $\Phi = \phi - \phi_d$.

The direction of the rupture propagation can be solved if both Δf and v are known. For a single source, the null position and spacing vary among receivers at different locations. Figure 4.25 shows the null positions in the event spectra for receivers whose locations are at different horizontal angles. The locations of the event and the receivers are shown in Figure 4.26(a). The peak frequency of this event is 280 Hz. Assuming $v = 1440$ m/s, the direction of the source propagation is shown as an arrow in Figure 4.26. For $v = 1000$ m/s or 1800 m/s, the direction of the propagation changes about 10 degrees. The data information and the calculation results are in Table 4.5. Note that the source direction agrees with all phones except D3.

The modeling of a propagating rupture by a phase-delay line array indicates that there is ambiguity in determining the characteristic frequency of the event, since the spectrum is multiplied by a sinc function which varies as observation position changes. Nulls in the event spectra are observed for all Type I events. With use of the above model, at least three inputs are needed to solve for source orientation, namely the peak frequency, null spacing, and propagation speed. Event data can only extract information of peak frequency and null spacing. Though the data are insufficient to extract the source parameters, the existence of nulls in event spectra indicates crack propagation.

4.4.4 Discussion of the Shear Fracture Model

Three aspects are addressed in this section:

1. Discussion on differences in S_{octo} for the two octopole groups.
2. Fitting of source data to the shear fracture model.
3. Plausibility of the shear fracture model.

As shown in the previous section, S_{octo} of close-in events (within about 300 m) lie between 1 and 100 kN/m for Type I events, and 1 and 1000 kN/m for Type II & III events. These calculations are consistent regardless of which octopole model is used. $S_{octo,lat1}$ for events further away (beyond about 500 m) lie in the same range as those close-in. Whether $S_{oct,lat2}$ is larger than $S_{octo,lat1}$ of far-away events, or far-away events are caused by Group I is an important issue to be verified. Since $S_{octo,lat2}$ increases with range, which increases the spread in $S_{octo,lat2}$, I suggest that far-away events are caused by ice motions modeled as octopoles of Group I. As for the close-in events, The verification of different octopoles is unnecessary since the results of S_{octo} are the same for the two groups.

To investigate further, dividing S_{octo} by $\langle B(\phi) \rangle_{rms}$ yields

$$S_o = \langle \frac{p_{peak} 2\pi r}{g(\theta)} \rangle = \frac{S_{octo}}{\langle B(\phi) \rangle} \quad (4.41)$$

Type	$B(\phi)$	Averaging Range of ϕ	$\langle B(\phi) \rangle$
$B_{lat1,1}$	$\sin(\phi) \cos(\phi)$	$(0, \pi/4)$	0.35
$B_{lat2,1}$	$\sin \phi$	$(0, \pi/2)$	0.71
$B_{lat2,2}$	$\cos \phi$	$(0, \pi/2)$	0.71

Table 4.6: Summary of all $\langle B(\phi) \rangle$

where S_o is the same as in Equation 4.25, and $\langle \rangle$ denotes the root-mean-square of the enclosed term. Table 4.6 shows the rms values for $B(\phi)$ and the range from $\phi_{min|B(\phi)|}$ to $\phi_{max|B(\phi)|}$ over which the rms is taken. The Brune stress drop obtained from S_o from Equation 4.27 then becomes

$$\sigma_B = S_o \frac{7}{16k^3 l_M} \left(\frac{\pi}{A} \right)^{3/2} \quad (4.42)$$

Assume that

$$\begin{aligned} f &= \frac{v}{L} \\ A &= \frac{v}{f} h \\ l_M &\sim h(\text{ice thickness}) = 2.5m \end{aligned}$$

where v is the fracture propagation speed. The fault plane is assumed to be on the $x - z$ or $y - z$ plane; h is in the z direction. Substitute $k = 2\pi f/c$ and the above into Equation 4.42, yields

$$\sigma_B = 5.9 \times 10^{-4} \left(\frac{c}{\sqrt{v}} \right)^3 f^{-3/2} S_o \quad (4.43)$$

The value for $\langle B(\phi) \rangle$ in Table 4.6 varies between 0.35 and 0.71, the same order of magnitude. Using $\langle B(\phi) \rangle = 0.59$ (the mean value), $S_{octo,lat1}$ in Equation 4.32, and $v = 1440$ m/s, σ_B becomes

$$\sigma_B = 2.7 \times 10^4 f^{-1 \pm 0.47} \quad (4.44)$$

From Equation 4.44, σ_B is 270 Pa at $f = 100$ Hz, and 27 Pa at $f = 1$ kHz. Figure 4.27 shows σ_B vs. peak frequency. Values of σ_B range predominantly from 10 to 1000Pa, which is far smaller than the ice fracture strength (about $0.2 \sim 1$ MPa).

Substituting $S_{octo,lat1}$ into Equation 4.41 and using $l_M = h = 2.5m$, the fracture moment in Eq 4.25 becomes

$$M_o = \frac{S_o}{k^3 l_M} = 4.1 \times 10^9 f^{-2.5} \quad (4.45)$$

At $f = 300$ Hz, M_o is 2630 N-m. Then the average slip dislocation in Equation 4.26 becomes

$$\bar{u} = \frac{M_o}{\mu A} = 3.8 \times 10^{-4} f^{-1.5} [m] \quad (4.46)$$

with $A = \frac{v}{f}h$, ice thickness $h = 2.5$ m, shear modulus $\mu = 3 \times 10^9$ Pa, and $v = 1440$ m/s. At 300 Hz, $\bar{u}$ is only 7×10^{-8} m which can hardly be observed. The estimated errors of the slopes for both M_o and $\bar{u}$ are 0.47, the same as that of $S_{octo,lat1}$ of Type I events.

Observations in seismology share the same fact that there is a 3 to 4-order-of-magnitude difference between the measured σ_B and the shear strength of the rock (in seismology), or the ice (my data). Brune et al. [6] reported there are earthquake events of high stress drop (about 100 bars) and low stress drop (about 0.1 bar). The low stress drop events (or partial stress drop events), like the ice events analyzed in this thesis, may occur when the fault locks itself right after the rupture passes, or when the stress release is not uniform and coherent over the whole fault plane. These phenomena may be due to asperities or barriers on the fault plane, or the complex geometry of the fault plane.

The above analyses focus only on the source strength of Type I events. However, the signature of the shear fracture can be any of Type I, II, V, or VI (Table 4.1). If the fracture progresses intermittently (stick-slip), Type II, V, and VI events are likely results of shear fractures.

The shear fracture in the ice floe is likely to occur under environmental forcing such as wave-induced stresses (Sec 4.2), tangential forcing due to floe shear (Sec 4.3.2), crushing force due to floe collisions (Sec 4.3.1), or wind and current stress during ridge breaking. Possible event signatures due to shear fractures are also observed,

and peak frequencies of the events ensure the fault plane dimensions in a reasonable range (the fault plane dimensions ranges from $1.4m \times 2.5m$ to $14.4m \times 2.5m$ at frequencies between 100 and 2000 Hz). Of course, the shear fracture is not limited by 2 kHz, the upper frequency limit of data. Source motions are confirmed by the existence of nulls in the event spectra, which is a characteristic of the shear fracture (see Table 4.1 and Sec. 4.4.3). Therefore, I conclude that *the shear fracture model is the most plausible source mechanism of the events in the MIZ at frequencies between 100 and 2000 Hz , and is a major contributor to the ambient noise in the MIZ at these frequencies.*

Figure 4.1: Time Series of a multi-pulse event which consists of 3 consecutive pulses of decreasing peak frequencies, namely 445, 124, and 60 Hz

Figure 4.1: Time Series of a multi-pulse event which consists of 3 consecutive pulses of decreasing peak frequencies, namely 345, 125, and 60 Hz

Figure 4.2: The spectrogram of Phone D3 of the event in Figure 4.1. Each contour level is 2dB (re $(Volt)^2/Hz$) increment. The peak frequencies of the three consecutive pulses are marked with ordering as 345 Hz, 125 HZ, and 60 Hz.

Figure 4.3: Geometry of a rigid circular plate tilting due to a force couple. The plate is attached to an infinite homogeneous thin plate. The ridge rotates due to environmental forcing.

Figure 4.4: The ratio of radial to tangential stress vs. r/l

Figure 4.5: A Vertical Dipole Model At the Sea Surface

Figure 4.6: Normalized Pressure Field of a Vertical Dipole vs Horizontal Range at two frequencies, namely 100 and 1000 Hz. The broken line is the 100 Hz curve, and the connected line is the 1000 Hz curve. The existence of ice is neglected. The sound speed profile for the water is shown in Fig. 1.3.

Figure 4.7: Dipole Force vs. Event Peak Frequency within Each Time Segment

Figure 4.8: Dipole Force vs. Event Peak Frequency within all Time Segments.

Figure 4.9: Dipole Force vs. Peak Frequency of Signature Type I to VI. The dash line is a least-square fit.

Figure 4.10: Dipole Force vs. Peak Frequency of Type I Events. The dash line is extrapolated from the central Arctic data for $f < 80$ Hz

Figure 4.11: Dipole Force vs. Decay Rate of Type II and III

Figure 4.12: Dipole Forces of Type I events from MIZ and Central Arctic compare with Ridge-unloading Model(Dyer89)

Figure 4.13: Strain at mid-Floe normalized by Surface Gravity Wave Magnitude versus Floe Length for Three Different Wave Periods, 12, 15, and 17 seconds. The Floe Thickness is 3 meters. (Wadhams88)

Figure 4.14: Floe Displacement before and after Breakup with wave period of 12 sec.(a) floe length = 100m,(b) 200m, and (c) 400m

Figure 4.15: Floe Length versus Peak Frequency of Unloading Motion When an Ice Floe Breaks.

Figure 4.16: Nine Types of Quadrupoles or Force Couples(Kim89). The axial arrangements are identical, as are the transverse (i,j); coordinate changes in each of these two groups distinguish one from another.

Figure 4.17: Spherical Coordinate System for the Octopole Model. The free surface is at $z = 0$.

Figure 4.18: Double Couple for Group I (Lateral Octopole I)

Figure 4.19: Two Possible Double Couples for Group II (Lateral Octopole II)

Figure 4.20: Lat1 Octopole Source Strength vs. Peak Frequency of Type I Events. The dashed line is a least-square fit (slope is 0.5). The error bars represents the range between maximum and minimum of octopole source strength inverted from each channels.

Figure 4.21: Lat2 Octopole Source Strength vs. Peak Frequency of Type I Events. The dashed line is a least-square fit (slope is -0.29). The error bars represents the range between maximum and minimum of octopole source strength inverted from each channels.

Figure 4.22: Lat1 Octopole Source Strength vs. Peak Frequency of Type II & III Events. The bar represents the range of the data. The dashed line is a least-square fit (slope is 1.4). The error bars represents the range between maximum and minimum of octopole source strength inverted from each channels.

Figure 4.23: Lat2 Octopole Source Strength vs. Peak Frequency of Type II & III Events. The bar represents the range of the data. The dashed line is a least-square fit (slope is 1.3). The error bars represents the range between maximum and minimum of octopole source strength inverted from each channels.

Figure 4.24: Comparison of Kim's result and octopole model. The receiver position: the horizontal range is 300 m, the depth is 60 m, and the horizontal angle is 30° . The fault plane is in the x-z plane. Both curves marked as Strike_Slip and Dip_Slip are from Figure 6.9 in (Kim89). The motion of the fault plane for strike-slip is in the x-direction (corresponds to OctopoleI), and for dip-slip is in the z-direction (corresponds to OctopoleII). Refraction is included by Kim with use of a mild SSP for the CA. My octopole model has no refraction (since refraction is not important at a horizontal range 300 m for the SSP of MIZEX84 in Figure 1.3).

Figure 4.25: Nulls of sinc function observed in event spectra

Figure 4.25: (a) Event and receiver locations of spectra shown in Fig. 4.23. The circles are the receivers, the cross is the event location, and the arrow is the hypodistal source orientation. (b) A series of the event in Fig. 4.23 and 4.25(a). The frequency range is 50 to 1000 Hz.

Figure 4.26: (a) Event and receiver locations of spectra shown in Fig. 4.25. The circles are the receivers, the cross is the event location, and the arrow is the hypothesized source orientation. (b) Time series of the event in Fig. 4.25 and 4.26(a). The frequency range is 60 to 1000 Hz

Figure 4.27: σ_B reduced from $S_{octo,lat1}$ in Fig. 4.20 vs. Peak Frequency. The dash line is a least-square fit (slope is -1)

Chapter 5

Conclusions

This thesis studies underwater ambient noise data from MIZEX84 to determine their acoustical origin. Analyses of noise events include classifying signatures, calculating spectra, locating events, estimating source strengths, and determining the plausible source mechanisms. This chapter discusses the major results in two parts – event statistics and source mechanisms.

5.1 Event Statistics

A total of 458 events, in four 40-minute time segments spanning over 20 hours, are analyzed over a region of about 1 km^2 in the Marginal Ice Zone. The frequency ranges between 20 and 2000 Hz during the 1st and 2nd segments and between 20 and 1000 Hz during the last two segments. The events are classified into six signature types (See Table 3.1). Each signature type is believed to be related to a different physical process (See Table 4.1). The signature types observed in the MIZ¹ are similar to those in the central Arctic.

Events tend to cluster in time and space. One third of the event clusters during the first 3 time segments recur at the dominant swell periods, between 10 to 18

¹After the completion of data analysis, I realize that all Type IV events classified previously in this data set should be identified as Type III events

seconds in the MIZ. However, during the 4th segment the cluster inter-arrival times fail to correlate with the swell periods, indicating that swells are important environmental factors during the first 3 segments but not during the 4th. The sea state measurements also confirm this finding – the swell reaches a maximum between 1800Z on J166 and 0400Z on J167, which encompasses the first 3 time segments. Then the swell dies down. Events also cluster in space during the first 3 segments but distribute more evenly during the 4th. These differences in the temporal and spatial distributions of events between the first 3 segments and the last may be due to changes in environmental forces or environmental conditions, such as swell peaks during the first three segments and ice concentration fluctuations from the 1st to the 4th segment. These two environmental factors of swell and ice concentration seem to dominate the distribution of events. The histograms of event inter-arrival times within each time segment show exponential behaviors which indicate that the probability of each individual event is identical and independent of other events, or event happening is a process without memory. The peaks within 0.01 to 10 s observed in the histograms of event inter-arrival time may be explained as fractures caused within one period of incoming swells, while the peaks within 9 to 20 s observed in the histograms of cluster inter-arrival time are due to fractures caused by several periods of incoming swells.

The chronological order and coherent locations of events in some clusters which inter-arrive at swell periods suggest crack propagation within the ice floe during the 1st segment. The cluster locations match the hypothesized floe boundary during the 2nd and 3rd segments, suggesting floe interactions to be a source of the events.

The event density varies with time, up to 10 times between the first and last time segment. The maximum event density is 0.58 per km^2 per hour per Hz at 300 Hz in the first segment. Event densities increase as f^1 at frequencies below 200 Hz, decrease as f^{-1} at frequencies above 300 Hz, and reach a maximum at frequencies between 200 and 300 Hz. The change in trend may indicate a transition in the source mechanism at frequencies below 200 Hz and above 300 Hz. Event

density is a threshold-dependent function. The threshold is the ambient noise level during event detection. The event densities reported should be understood as only a portion of the actual event densities.

5.2 Source Mechanisms

Four source mechanisms are reviewed, which include the ridge-unloading model, floe breakup due to surface gravity waves, floe interactions, and shear fractures which co-exist with floe-interactions and precede the ridge-unloading model and floe breakup.

The shear fracture model is the most plausible source mechanism at frequencies above 100 Hz. The unloading motion of the deformed ice plate, which occurs right after the plate breaks, is the possible source at frequencies between 10 and 100 Hz (which includes the ridge-unloading model and the floe breakup model due to surface gravity waves). Mid-frequency pulses (125 Hz and 345 Hz) taking place before low-frequency pulses (60 Hz) as shown in Figure 4.2 are important findings, because they confirm that shear fractures precede the unloading motion of a deformed ice plate when the plate breaks. The next two sections present discussions and results on these two source mechanisms.

5.2.1 Unloading motion (10 - 100 Hz)

The unloading motion of a deformed ice sheet, after the plate breaks, is shown as a likely source mechanism at frequencies below 100 Hz. The ice deforms under the stresses caused by surface gravity wave, wind, and current. Many fractures and cracks in the ice precede this. A vertical dipole is used to represent the volume change in the unloading process. The dipole forces of the Type I events agree well with the extrapolated values from the previous results for the central Arctic. However, the ice stresses corresponding to the dipole forces are far less than the ice strength. The creep behavior of the ice at low strain rate may be the cause

of this low stress result [14], although it seems unlikely for the MIZ in contrast to the central Arctic for which it was first invoked (Since the loading rate of the deformation of ice plate is much faster in the MIZ than that of the CA).

5.2.2 Shear Fracture Model (100 Hz and above)

Shear fracture is the most plausible source mechanism at frequencies above 100 Hz. An octopole, obtained from a double couple and its image, is used to represent the shear fracture at a fault plane. The double couple moment (or fracture moment, or seismic moment used in seismology) of Type I events ranges between 10^2 and 10^5 N-m; its frequency dependence is $f^{-2.5 \pm 0.46}$, where -2.5 is resulted from least square fit method. This least-square-fit is considered as a preliminary effort to define the relationship between the moment and the peak frequency; the error involved will not deter the conclusion on plausibility of shear fracture model.

The Brune stress drop of the observed shear fractures is 3 to 4 orders of magnitude smaller than the shear strength of the ice. This low stress drop, also observed in seismology, may be the result of fault locking, or the non-uniform distribution of the stress release on the whole fault plane, caused by asperities on the fault plane.

Ice existence is neglected in the octopole model. This assumption is examined by comparing the octopole model with Kim's model [29] which placed a double couple at the fault plane embedded in an infinite ice plate. The source strength of the Group I octopole is ten times (20 dB) less than that of Kim's model at frequencies above 300 Hz. The fracture moment, or the stress drop calculated based on the octopole model, should be multiplied by 10 at frequencies above 300 Hz; the ice therefore cannot be neglected since the wavelength is comparable with the ice thickness at these higher frequencies.

Fracture propagation is one of the characteristics of shear fractures, which are modeled by a phase-tapered array. The spectra of received signals are multiplied by a sinc function with parameters such as the direction of fracture propagation,

the propagation speed, and the fault plane dimension. The fault plane dimension is related to the peak frequencies observed in event spectra. However, the data are still not sufficient to extract the other two parameters, namely the propagation speed and the source orientation. Nevertheless, the nulls observed in the spectra of Type I events, which are the characteristics of a sinc function, indicate fracture propagation.

5.3 Recommendations

Several things can be done for future research on experiment setup, ice model, and event statistics:

- Event statistics could be further investigated with more direct environmental measurements than those available in MIZEX84. Swell measurements could be obtained by installing accelerometers attached to buoys in the ice floe in which the hydrophones are attached [40]. Ice concentration or floe distribution could be measured by taking aerial photographs over the area where phones are located. This area is about 1 km^2 in the case of this data set (since the effective detection area of the arrays used is about 1 km^2 . See Section 3.4).
- Further work in modeling the relationships between the event density and the environmental forcings is useful such that the observed event densities can be linked to ambient noise levels and environmental forces.
- The analysis of the source strength of shear fractures focuses only on Type I events. Other signature types such as Type II, V, and VI can be caused by intermittent fracture processes. Further research should be done on events of these signature types.
- Ice effect should be included in shear fracture model. The numerical model developed by Kim [29] can be used to invert source strength from the data.

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Appendix A

Data Analysis

A.1 Digitized Data Format

The digital data acquisition system used in MIZEX84 is described in Prada et al. [52]. Figure A.1 shows the characteristics of the built-in low-pass and high-pass filters. The on site digitization had a 1 kHz sampling frequency, which resulted in a 250 Hz cut-off. The data were stored on magnetic tapes. The tape format is shown in Figure A.2.

I use data recorded in analog form. The analog recording tape speed varies among these four periods. I use the same digitization system as described above except at a higher sampling frequency, by slowing down the speed of the analog tape being digitized. Table A.1 shows the analog tape speed and the sampling frequency for each segment.

The data recording was not continuous for both digitization and analog recording. There were also reflection, reverberation, and refraction experiments which fired shots in intervals. The transmission experiment at times transmitted tones in the frequency range of present interest. Thus only four 40 minute segments out of a day were selected for analysis.

Seg. #	Tape #	Tape Speed	Sampling Freq.
I	A7	7.50 ips	8 kHz
II	A8	7.50 ips	8 kHz
III	A9	3.75 ips	4 kHz
IV	A10	3.75 ips	4 kHz

Table A.1: Digitization information for the four segments

A.2 Waterfall Plot Technique

A waterfall plot, like a seismograph, contains time series for all recording channels. Thus outstanding features on all channels can be simultaneously detected by eye.

The frequency range of interest and duration of events are the two major considerations of waterfall plots. Time series are plotted with a high pass filter of 20 Hz due to cable strum contamination at lower frequencies. With durations of most events less than 0.1 s, the plotting time is chosen to be larger than 0.1 s, keeping the same order of time scale. Figure 2.2 presents data from 10 to 1000 Hz on 12 recording channels over a total plotting time of 0.15 s, divided into five 0.3 s segments. Events are clearly seen in the plot.

A.3 Time Delay Error Estimation

Knapp et al. [30] model signals from a remote source received by two sensors as:

$$x_1(t) = s(t) + n_1(t)$$

$$x_2(t) = \alpha s(t + D) + n_2(t)$$

where

$x_i(t)$ = time series received by the i th sensor,

$s(t)$ = signal emitted by the source,

$n_i(t)$ = noise at the i th receiver,

α = difference in transmission loss between two receivers, and

D = time delay of the received signal at 2nd receiver.

The time delay estimator is

$$R_{y_1 y_2}^{(g)}(\tau) = \int_{-\infty}^{\infty} \psi_g(f) G_{x_1 x_2}(f) e^{j2\pi f\tau} df \quad (\text{A.1})$$

where

$y_i(t)$ = filtered signal of $x_i(t)$,

$\psi_g(f) = H_1(f)H_2^*(f)$,

$H_i(f)$ = complex frequency response of the filter for $x_i(t)$,

$H_i^*(f)$ = complex conjugate of $H_i(f)$, and

$G_{x_1 x_2}(f)$ = cross power spectral density function of $x_1(t)$ and $x_2(t)$ of finite observation.

Throughout the analysis for the time delay error,

$$\psi_g(f) = \begin{cases} 1, & f_{min} \leq f \leq f_{max} \\ 0, & \text{otherwise} \end{cases}$$

where f_{min} and f_{max} are the lower and upper limit of an event frequency range. The processor of time delay determination is illustrated in Figure A.3.

The variance of time delay obtained by the above equation is

$$Var[D] = \frac{\int_{-\infty}^{\infty} |\psi(f)|^2 (2\pi f)^2 G_{x_1 x_1}(f) G_{x_2 x_2}(f) [1 - |\gamma(f)|^2] df}{T [\int_{-\infty}^{\infty} (2\pi f)^2 |G_{x_1 x_2}(f)| |\psi(f)| df]^2}$$

where

$$\gamma(f) \triangleq \frac{G_{x_1 x_2}(f)}{\sqrt{G_{x_1 x_1}(f) G_{x_2 x_2}(f)}}$$

is the coherence estimate of $x_1(t)$ and $x_2(t)$.

Several examples of calculations of variance are shown as follows:

Example 1:

The noise at receivers is assumed to be incoherent and identical in property. The power spectra of the signal and the noise are

$$G_{ss}(f) = \begin{cases} S_o, & f_{min} \leq f \leq f_{maz} \\ 0, & \text{otherwise} \end{cases}$$

$$G_{nn}(f) = \begin{cases} N_o \left(\frac{f_{min}}{f}\right)^m, & f_{min} \leq f \leq f_{maz} \\ 0, & \text{otherwise} \end{cases}$$

and

$$SNR = \frac{S_o}{N_o} \frac{(1-m)(f_{maz} - f_{min})}{\left(\frac{f_{min}}{f_{maz}}\right)^m f_{maz} - f_{min}}$$

The above equations are shown in Figure A.4. With $f_{min} = 200Hz$, $f_{maz} = 2kHz$, $S_o = N_o$, $m = 2$, the time delay variance is

$$Var[D] = \frac{2.57\alpha^2 + 2.83}{T\alpha^2} \times 10^{-12}$$

For a typical event, the range of T is between 0.01 s and 1 s and α of different channel is between 0.1 and 0.9. Thus the time delay error ($= \sqrt{Var[D]}$) is about 3 to 53 μs .

Example 2:

Let $G_{ss}(f)$ be of the same form as $G_{nn}(f)$ except with different power law, i.e.

$$G_{ss}(f) = \begin{cases} S_o \left(\frac{f_{min}}{f}\right)^p, & f_{min} \leq f \leq f_{maz} \\ 0, & \text{otherwise} \end{cases}$$

With $G_{nn}(f)$ the same as in Example 1 and $p = 1$,

$$Var[D] = \frac{2.98\alpha^2 + 4.14}{T\alpha^2} \times 10^{-12}$$

For $T = 0.01$ s to 1 s and $\alpha = 0.1$ to 0.9, the time delay error is between 2.7 and 204.2 μs . If m remains the same, $p = 4$ and $S_o = 5N_o$, then

$$Var[D] = \frac{5.9 \times 10^{-14}\alpha^2 + 1.3 \times 10^{-9}}{T\alpha^2}$$

Phone #	r_i [m]	$\sin \theta_i$
D3	100.3	0.60
D5	68.3	0.88
D11	76.4	0.79
D12	65.1	0.92
D18	63.1	0.95

Table A.2: Phone information for Example #3

For $T = 0.01$ s to 1 s and $\alpha = 0.1$ to 0.9, the time delay error is between 30 and 939 μ s.

Example 3:

Now assume the source as a vertical dipole, then $x_i(t)$ becomes

$$x_i(t) = \frac{\sin \theta_i}{4\pi r_i c} \frac{\partial F(t + D_i)}{\partial t} + n_i(t)$$

and the cross power spectra are

$$G_{x_i x_i} = \left(\frac{\sin \theta_i f}{2r_i c} \right)^2 |\mathcal{F}(f)|^2 + G_{nn}(f)$$

$$G_{x_1 x_2} = \frac{\sin \theta_1 \sin \theta_2 f^2}{4r_1 r_2 c^2} |\mathcal{F}(f)|^2$$

with the similar assumptions as before that the noise are identical and incoherent at two sensors. Let

$$|\mathcal{F}(f)|^2 = \begin{cases} F_o \left(\frac{f_{min}}{f} \right)^p, & f_{min} \leq f \leq f_{max} \\ 0, & \text{otherwise} \end{cases}$$

and $G_{nn}(f)$ remains to be the same.

Consider an actual event as an example. Figure A.5 shows the time series. The frequency range of the event is between 200 and 2000 Hz with a duration of 0.05 s. The relevant information for all five phones are listed in Table A.2.

The relative levels of noise and the dipole force can be solved by comparing the event power spectrum and the ambient noise power spectrum. For Phone D18,

$$|X(f_{min})|^2 = 1.093 \times 10^{-7} F_o + N_o$$

with the event power spectrum about 6 dB higher than ambient noise, yields $|X(f_{min})|^2 = 2N_o$. Solving for F_o , gives

$$F_o = 9.15 \times 10^6 N_o$$

Hence let $F_o = 10^7 N_o$, $m = 2$, and $p = 3$, the time delay error (σ_r) is 0.0048 s for Phones D12 and D18 and is 0.009 s for Phones D12 and D3.

A.4 Range Error Estimation

The range of an event from the source to the receiver is determined by two inputs, namely time delays and receiver positions. So the range error σ_r depends on the a) time delay error σ_r , b) travel time error σ_t , and c) receiver positions error σ_l . The value for σ_r is 5×10^{-3} s (see appendix A.3) and σ_t is 0.004 s (assuming a direct path instead of a refractive path based on the sound speed profile in Figure 1.3). The error in phone positions depends on two issues. One is the estimation error resulted from the Sensor Tracking System, which is about 1 m. The other is the uncertainty of actual phone position due to swinging motion caused by underwater currents.

The drift phone configuration is shown in Figure A.6. The 62 m cable was frozen at the ice top and was tied with a 12 lb ballast at the bottom. Assuming constant differential current along the cable length (since the phone and cable drifted with ice floe), the drag on the cable due to drift is [3]

$$R = \frac{1}{2} \rho_w C_{DN} d V^2$$

where

ρ_w = density of sea water (= 2 lbf/ft),

V [knot]:	0.1	0.2	0.4	1.0
x [ft]:	1.9	7.5	29.0	117.0

Table A.3: Current speed and cable excursion

C_{DN} = drag coefficient ($\cong 1.4$ for hairy cable below critical Reynold's number),

d = cable diameter (1/16 in), and

V = differential current speed.

The force equilibrium is shown in Figure A.6. Assuming a) the top end of the cable is in vertical direction, b) the angle at the lower end is ϕ_b , and c) the cables are neutrally buoyant, the tension in the cable is

$$T_o = W \times \csc \phi_b = S \times R \tan \phi_b$$

The cable excursion is

$$x = \frac{T_o}{R} (\csc \phi_b - 1)$$

where S is the cable length and W is the ballast weight.

Table A.3 shows values for the differential current speed V and the cable excursion x ,

Note that x is proportional to V^2 when V is small. For the position error, I use $V = 0.2$ knots which yields 7.5 ft of cable excursion. The total error in the phone position is

$$\sigma_l = \sqrt{7.5^2 + 1^2} = 7.6 \text{ ft} = 2.3 \text{ m}$$

So range error is

$$\begin{aligned} \sigma_r &= \sqrt{[c \times (\sigma_r + \sigma_l)]^2 + \sigma_l^2} \\ &= \sqrt{(13.0)^2 + (2.3)^2} \\ &= 13.2 \text{ m} \end{aligned}$$

In the analysis for locating events (see Section 2.4), I actually use 10 m for σ_r .

A.5 Lloyd Mirror Effect Correction

The Lloyd mirror effect is the interference of the image created by free surface ([9] or other text books about acoustics or optics). Figure A.7 shows the source, its image and the receiver positions. The source depth is d , and the angle below the free surface is θ , then the pressure at the receiver is

$$p(\underline{r}, t) = \frac{f(\theta)B(\phi)}{4\pi r} S(t - r/c) [-2i \sin(kd \sin \theta) + \frac{d \sin \theta}{d} (2 \cos(kd \sin \theta))] \quad (\text{A.2})$$

where $f(\theta)$ and $B(\phi)$ are the radiation patterns of the source $S(t)$ in an unbounded media, r is the distance from the mid point of the source and the image to the receiver. For far field approximation, the first term is retained. Then

$$p(\underline{r}, t) = \begin{cases} -i \frac{2 \sin(kd \sin \theta)}{4\pi r} S(t - r/c), & kd \sin \theta \gg 1 \\ -i \frac{2kd \sin \theta}{4\pi r} S(t - r/c), & kd \sin \theta \ll 1 \end{cases} \quad (\text{A.3})$$

So the mirror effect correction is $\text{Sinc}(kd \sin \theta)$ such that

$$p_{w/mirror} = \frac{\sin(kd \sin \theta)}{kd \sin \theta} \times p_{w/omirror} \quad (\text{A.4})$$

(a)

Figure A.1: (a) Low-pass Filter, (b) High-pass Filter of Digital Data Acquisition System

Figure A.2: Digital Tape Format

Figure A.3: Time Delay Processor[30]

a) Cross Power Spectra of Example #1

b) Cross Power Spectra of Example #2

Figure A.4: Example # 1

Figure A.5: Time Series of Example #3

Figure A.6: The Configuration and Force Equilibrium Diagram of Drift Phone Cable

Figure A.7: Geometry of the source and its image, and the receiver